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AN EMPIRICAL STUDY OF VERTICAL VELOCITIES IN THE LOWER STRATOSPHERE*

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ABSTRACT

Vertical velocities have been computed for the lower stratosphere for two independent winter periods, by employing a form of the adiabatic method. The regions studied were in both cases outside the polar vortex. The flow pattern was divided into stationary (long-wave) and transient (short-wave) components. The vertical velocity pattern associated with the stationary long wave is precisely that described by Kochanski [3]; i.e., the air rises in moving from warm troughs to cold ridges. The pattern associated with the short waves is more complicated. There is a maximum of warm advection in the vicinity of short-wave ridges, and cold advection near troughs. Local temperature changes, however, very nearly compensate the advection, with the net result that in the mean the vertical velocities associated with short-wave patterns are small, but tend to be positive near ridges and negative near troughs. Superimposition of the short and long wave, however, can lead to any conceivable combination of signs of advection, local temperature change, vertical velocity, and position with respect to ridge or trough.

The single parameter which is most useful in specifying the vertical velocity is the temperature advection.

1. INTRODUCTION

In the middle 1940's, spurred by recent developments in atmospheric sounding techniques, meteorologists began to investigate in considerable detail the nature of the vertical motions in the lower and middle troposphere and their relation to the horizontal flow. Now, in the late 1950's, as further improvements have increased the density and reliability of data in the lower stratosphere, it is feasible to extend our investigations to those levels. The present study is offered as an early step in that direction. A hypothesis will be formulated concerning the manner in which the vertical velocities are related to the troughs and ridges in the lower stratosphere, in the hope that this may serve as a basis for further studies.

The earliest attempt at an empirical evaluation of the vertical velocities in the lower stratosphere was that of Kochanski in 1951 (cf. [3]). Mean soundings at troughs and ridges were compared, and vertical velocities were computed on the assumption of adiabatic motion by computing the time it would take for parcels of air to move along a contour from ridge to trough. Average vertical velocities of the order of 3 cm. sec.$^{-1}$ were found near the tropopause, diminishing to almost negligible amounts just above 100 mb. Brockmeyer [1] studied vertical velocities in the lower stratosphere for the period January 1–10, 1956 by means of the usual adiabatic method. His results were quite similar to those of Kochanski; i.e., values of vertical velocity ranging up to 3 to 4 cm. sec.$^{-1}$ at 200 mb. with smaller values at higher levels, and, in general, subsidence from ridge to trough and rising motion from trough to ridge.

Craig and Hering [2], in a study of the sudden warming in the polar stratosphere in January 1957, estimated that to account for changes in temperature along a trajectory, assuming adiabatic processes, vertical velocities of about 5 cm. sec.$^{-1}$ were required at 25 mb. This, of

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course, was a somewhat anomalous situation. Teweles [4], in discussing the same situation, estimated average rates of subsidence of 1 to 3 cm. sec.$^{-1}$ over somewhat longer paths.

2. METHOD

The technique chosen for computing vertical velocities is a modified version of the adiabatic method. This method is particularly well suited to the stratosphere because of the large stability in that region. The basis of the adiabatic method is the equation:

$$w=-(\mathbf{V}\cdot\nabla T+\partial T/\partial t)/(\gamma_{ad}-\gamma). \quad (1)$$

The assumption is made that the horizontal advection of temperature ($\mathbf{V}\cdot\nabla T$) is very nearly equal to constant pressure advection.

To simplify the computations it is also assumed that the lapse rate (γ) may be neglected in the stratosphere, in comparison with the adiabatic lapse rate (γ_{ad}). This is not likely to be in error, in the regions studied, by more than 15 or 20 percent. Finally, then, if the vertical velocity (w) is measured in units of cm. sec.$^{-1}$, and if the units of advection and local temperature change are °K.(12 hr.)$^{-1}$, equation (1) may be rewritten

$$w=-0.24(\mathbf{V}\cdot\nabla T+\partial T/\partial t). \quad (2)$$

In most previous applications the right hand side of equation (2) has been evaluated by estimating temperature changes along trajectories. This is unsatisfactory for two reasons. First, there is the difficulty of constructing trajectories from synoptic maps; second, the vertical velocities computed are then averages both in time and space over the length of the trajectory. Since a parcel of air may travel hundreds or even thousands of kilometers in 12 hours, this leads to considerable ambiguity in interpreting the results with respect to the synoptic pattern.

By evaluating the two terms on the right hand side of equation (2) separately, a method is developed by which adiabatic vertical velocities, although averages over 12-hour time intervals, refer to specific areas, which, in the present case, are 10° latitude by 10° longitude. There is no longer any question as to where, along a trajectory, the vertical velocities occurred.

The basic grid is indicated in figure 1. The input information was the temperature at every point in the 5° x 5° grid (numbers in fig. 1) and mean winds within every 5° box (indicated by letters). The temperatures were estimated to the nearest ½° K. on the basis of an isothermal analysis made for every 2° K. The vector winds were estimated subjectively to the nearest 5 knots and 10°, from observed winds wherever possible, but supplemented by geostrophic winds when necessary.

The mean temperature at a point, such as point 5 in figure 1, is defined as:

$$\overline{T}_5=\frac{1}{16}(T_1+2T_2+T_3+2T_4+4T_5+2T_6+T_7+2T_8+T_9).$$

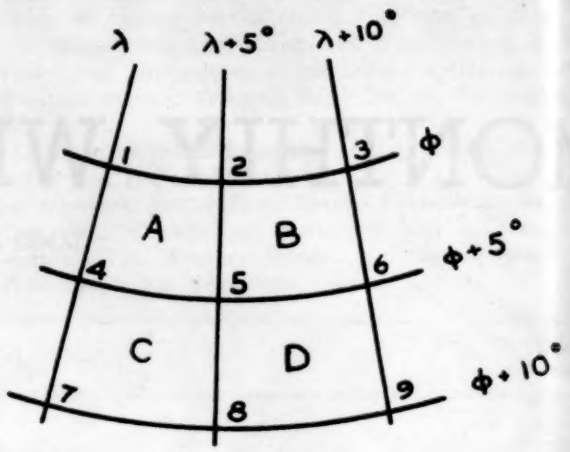

FIGURE 1.—Grid employed in vertical velocity computations.

The mean advection at the same point is defined by:

$$(\overline{\mathbf{V}\cdot\nabla T})_5=\frac{1}{4}[(\mathbf{V}\cdot\nabla T)_A+(\mathbf{V}\cdot\nabla T)_B+(\mathbf{V}\cdot\nabla T)_C+(\mathbf{V}\cdot\nabla T)_D]$$

where

$$(\mathbf{V}\cdot\nabla T)_A=\frac{Cu_A}{\cos(\varphi+2\frac{1}{2}°)}[(T_2-T_1)+(T_5-T_4)]+Cv_A[(T_1-T_4)+(T_2-T_3)].$$

Here u_A and v_A are, respectively, the west-east and south-north wind components corresponding to the mean vector wind in box A. C is a proportionality factor which involves the grid size and the conversion of units.

Averages over time (12 hours) of advection and temperature change were accomplished by averaging the advection fields on two successive maps and subtracting the temperature fields. In other words, if the times of successive maps are t_0 and t_0+12,

$$\left(\overline{\frac{\partial T}{\partial t}}\right)_5^{(t_0+6)}=\overline{T}_5^{(t_0+12)}-\overline{T}_5^{(t_0)}$$

and

$$(\overline{\overline{\mathbf{V}\cdot\nabla T}})_5^{(t_0+6)}=\frac{1}{2}[(\mathbf{V}\cdot\nabla T)_5^{(t_0)}+(\mathbf{V}\cdot\nabla T)_5^{(t_0+12)}].$$

3. SOME EXAMPLES OF THE RESULTS

Computations of advection and local temperature change in the lower stratosphere (200, 150, and 100 mb.) were made for two winter periods, December 17–23, 1953, and January 1–10, 1956. The latter period had been studied by Brockmeyer [1] by means of the trajectory technique.

Figures 2–5 are examples of some of the results of the study. The contour patterns at 150 mb. for December 17, 1953 and for January 4, 1956 are shown in figures 2 and 3 along with the corresponding patterns of vertical velocity. Figures 4 and 5 contain latitudinal cross sec-

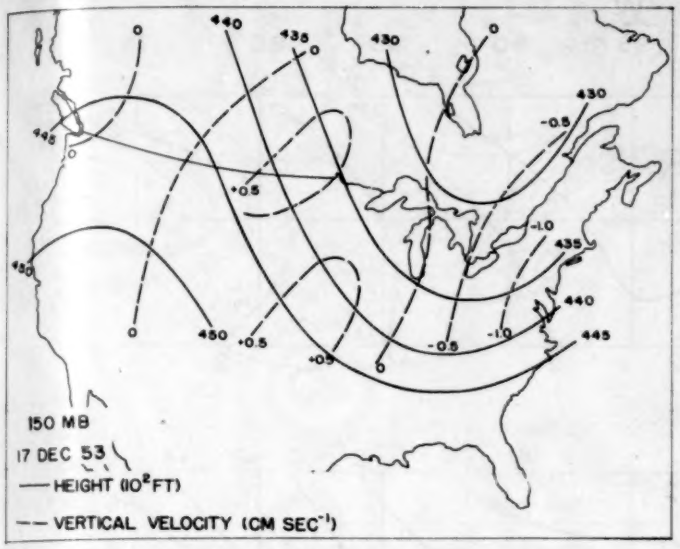

FIGURE 2.—Contours and vertical velocities at 150 mb. on December 17, 1953.

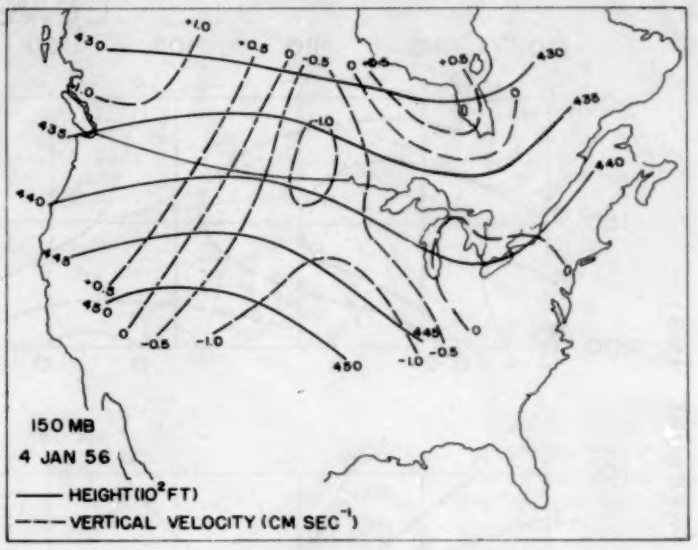

FIGURE 3.—Contours and vertical velocities at 150 mb. on January 4, 1956.

tions on the same dates. In all four figures, the vertical velocities are averages of two successive sets of computations; i.e., they are 24-hour averages.

On December 17 (fig. 2) there was a pronounced ridge near 115° W. longitude and a trough near 75° W. Between these two meridians was a zone of positive vertical velocities above 200 mb. The magnitude of the vertical velocities was not large but a smooth pattern is indicated and it persisted for the next several periods. The evidence from the cross section (fig. 4) is that there was slight horizontal convergence in this region. Although the vertical velocities were still positive (or zero) even at 200 mb., the pattern suggests that within the troposphere the air was probably subsiding as it moved from ridge to trough. Thus, it appears there was a reversal of the sign of the vertical velocity between the troposphere and the lower stratosphere.

The case of January 4 (fig. 3) tells a different story. Here there was a ridge near 105° W. longitude and a trough near 80° W. In this case, however, the region between these meridians was one of pronounced subsidence. The cross section (fig. 5) shows that the subsidence increased downward. There was no change of sign at the tropopause.

Meteorologists have learned to expect subsidence and convergence in the upper troposphere in association with flow of air from ridge to trough. Neither of the cases presented here, in spite of their differences, conflicts with this concept. In both cases, the indicated sign of the convergence in the lower stratosphere is in agreement with that which we would expect in the upper troposphere. In one case this convergence resulted in decreasing subsidence with increasing height; in the other the vertical gradient of vertical velocity ($\partial w/\partial z$) appears to have been of such a magnitude that there was a reversal of the sign of the vertical velocity between the upper troposphere and the lower stratosphere, and the rising motion in the lower stratosphere increased with height.

We are confronted, then, with one case in which the stratospheric pattern of vertical velocities was that with which we are familiar in the troposphere; i.e., subsidence in air moving with a northerly component; and another in which the sign of the vertical velocity was reversed. It should be pointed out here that the two examples presented here are not extreme cases, although they were chosen to illustrate the wide range of vertical velocity patterns. It should also be pointed out that all possible combinations of vertical velocity and meridional flow were encountered, particularly during the period of December 1953.

4. THE LONG-WAVE PATTERN

In an attempt to create some understanding out of the seeming disorder apparent on the daily maps, each sequence was separated into a time-mean component and deviations from the mean. The means were computed by averaging the values at each grid point, and here are identified with the long-wave systems which are indeed very nearly stationary. The deviations from the means are identified with the short-wave systems. In this section we shall discuss only the long-wave (mean) components.

The map in figure 6 illustrates the mean contour, isotherm, and advection fields at 150 mb. for the period December 17–23, 1953. (The mean advection is determined not by $\overline{\mathbf{V}}\cdot\overline{\nabla T}$, but rather as the average of all the fields of advection ($\overline{\mathbf{V}\cdot\nabla T}$). Any correlation between the deviations of $\mathbf{V}$ and ∇T from their means will cause differences between these two types of averages. Thus, the field of advection in figure 6 (and also in fig. 7) need not agree with that which would be given by the contours and isotherms on the maps). Over this period there was a slight

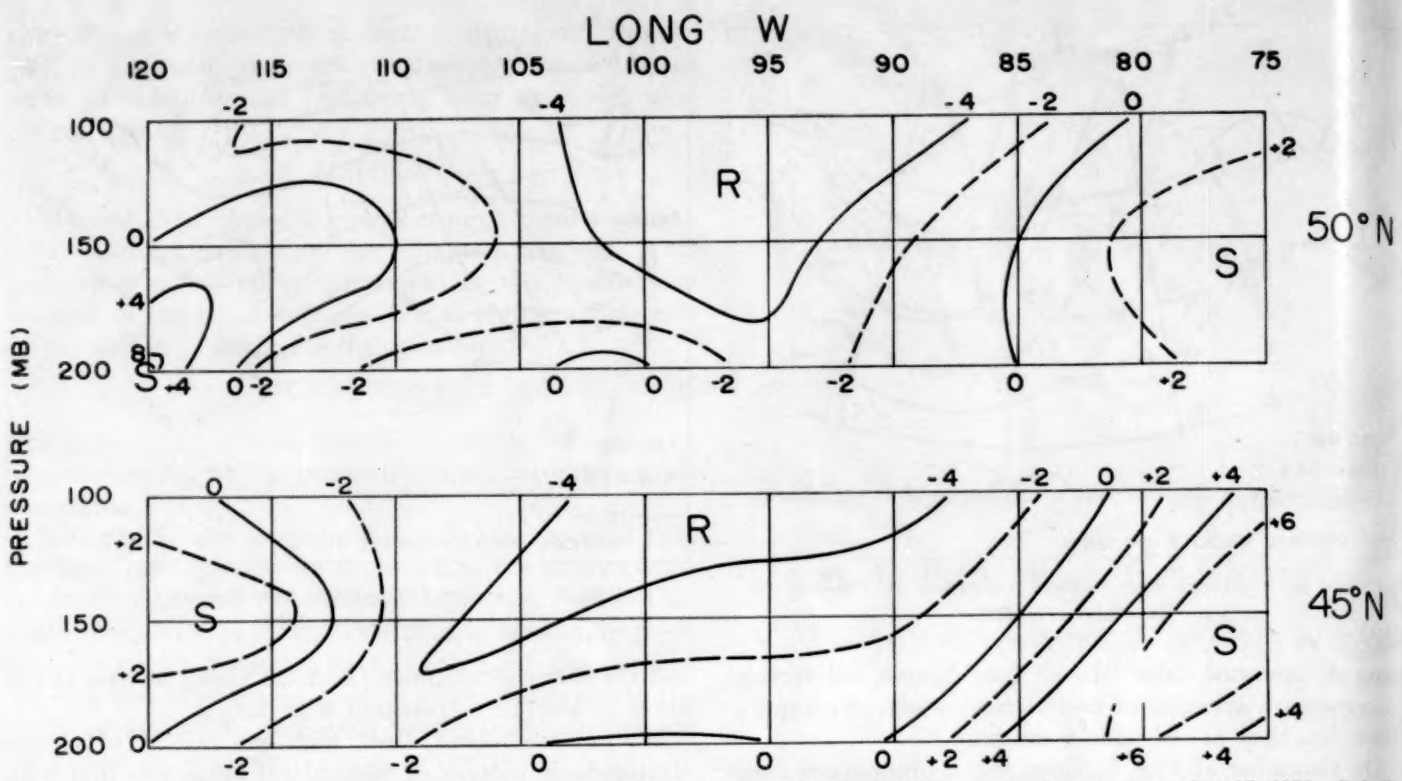

FIGURE 4.—Vertical velocity cross sections in the lower stratosphere along latitudes 45° N. and 50° N. on December 17, 1953. R indicates rising motion, S subsidence. Isopleths are labeled in units of mb. (12 hr.)$^{-1}$. This choice of units, plus a linear pressure scale as the abscissa, makes it possible to interpret the vertical velocity directly in terms of convergence or divergence.

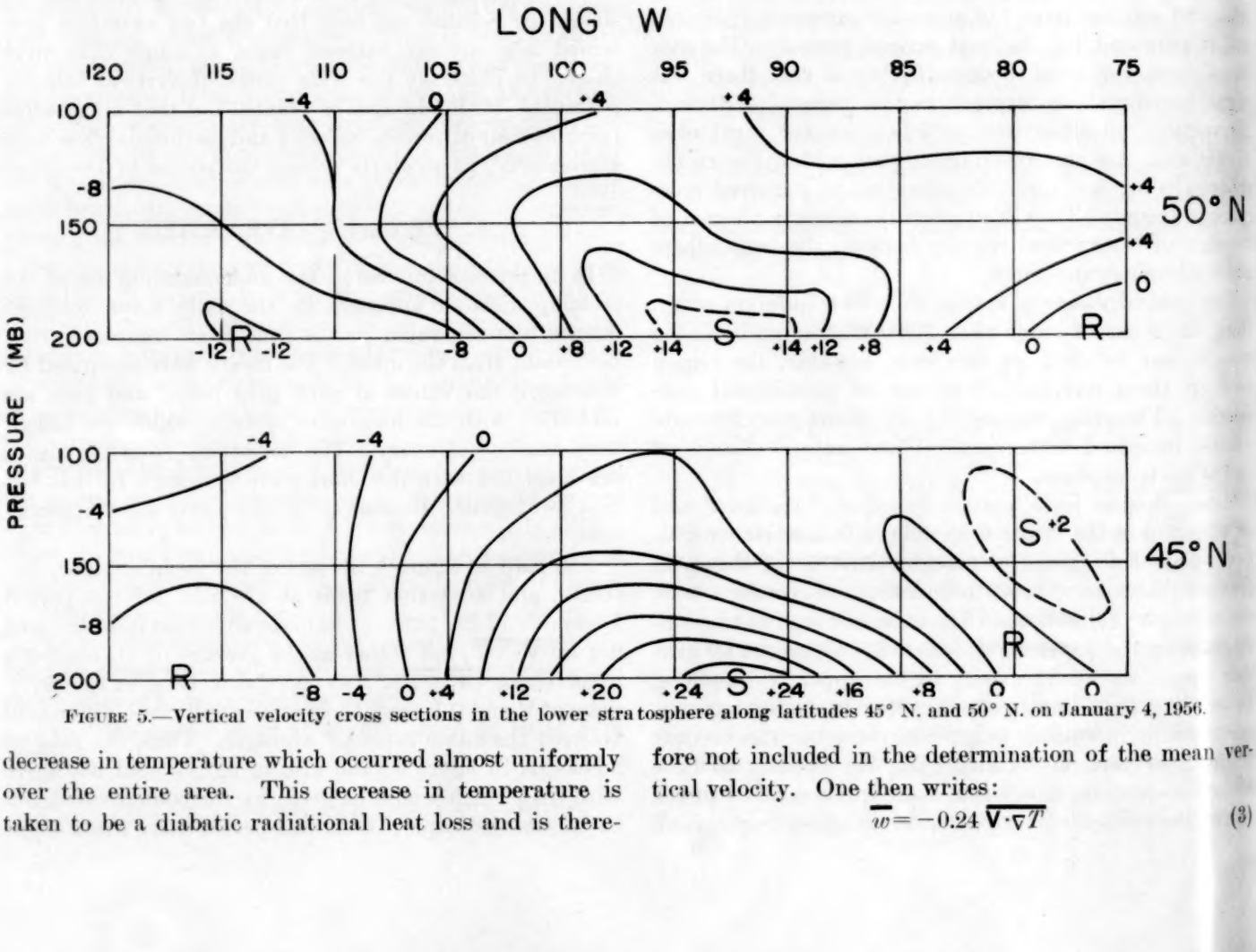

FIGURE 5.—Vertical velocity cross sections in the lower stratosphere along latitudes 45° N. and 50° N. on January 4, 1956.

decrease in temperature which occurred almost uniformly over the entire area. This decrease in temperature is taken to be a diabatic radiational heat loss and is therefore not included in the determination of the mean vertical velocity. One then writes:

$$\overline{w} = -0.24\,\overline{\mathbf{V}\cdot\nabla T} \tag{3}$$

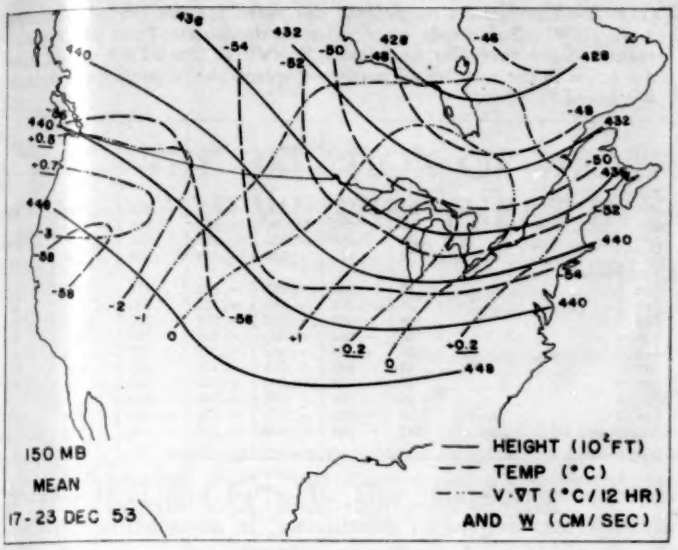

FIGURE 6.—Mean 150-mb. map for the period December 17–23, 1953. The isopleths of mean temperature advection are labeled in °K. (12 hr.)$^{-1}$ and also as mean vertical velocities in cm. sec.$^{-1}$ according to equation (3). The underlined values are the vertical velocities.

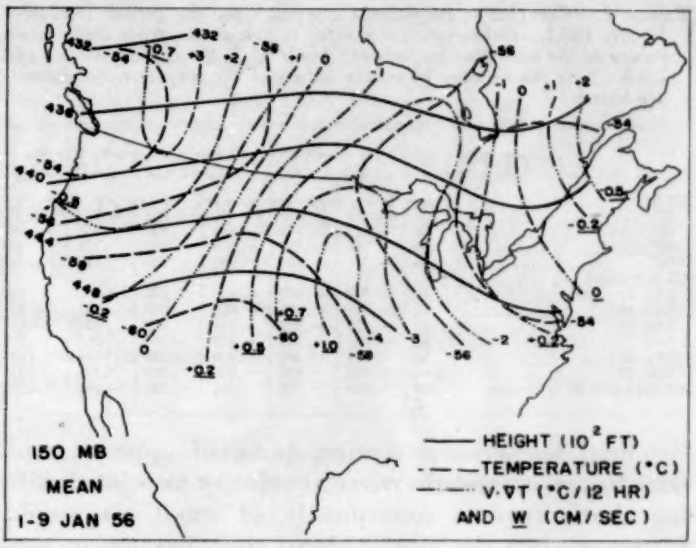

FIGURE 7.—Mean 150-mb. map for the period January 1–10, 1956. Isopleths are labeled as in figure 6.

Certainly, if the long waves are indeed stationary and in a steady state, there should be no local temperature changes which could be associated with them. Thus, the isopleths of advection in figure 6 are labeled at one end in units of °K. (12 hr.)$^{-1}$ and at the other in units of cm.sec.$^{-1}$.

The mean map for the period January 1–9, 1956 is given in figure 7. In this case, it should be pointed out, the averages of the local temperature changes, although everywhere small (mostly less than ¼° K. per 12 hours), were not uniformly negative and did exhibit some semblance of a pattern. This is due to a combination of effects: the fact that an integral number of short waves need not have passed a given location during the period; non-uniform radiational effects caused by differences in the temperature and nature of the earth's surface and of the air itself; also, a slight change of the long-wave position may have occurred. Nevertheless, the local temperature changes were sufficiently small that they are neglected in figure 7, and equation (3) was employed directly as in figure 6.

The fact that the average local temperature changes were so small and uniform offers a considerable degree of "a posteriori" justification for identifying the mean fields with the atmospheric long waves. If the long waves had moved significantly during the periods studied they would have carried with them their temperature fields, and this movement would have shown up in the average temperature derivatives. Examination of the daily maps of the deviations of the contour fields from their means (not shown here) also supports the contention that the long waves were very nearly stationary during these periods.

Both situations exhibit certain fundamental features which we have come to recognize as characteristic of the lower stratosphere outside the polar vortex, namely, warm troughs and cold ridges, with the isotherms showing greater amplitude than the streamlines. This corresponds, as Kochanski pointed out, to air descending from ridge to trough and ascending from trough to ridge. Kochanski's results indicated an average vertical velocity of 1.4 cm. sec.$^{-1}$ near the point of inflection at 140 mb., a value which is in reasonable agreement with those shown in figures 6 and 7.

5. SHORT-WAVE COMPONENTS

One may obtain an array of deviations from their means of $\mathbf{V}\cdot\nabla T, \partial T/\partial t, w$, and h (the height of the 150-mb. surface) by subtracting from the daily mean values the appropriate means as shown in figures 6 and 7 (and also the mean values of $\partial T/\partial t$). These deviations are taken as representing the effects on each of the parameters of the migratory short-wave disturbances. To study the interrelations of the various terms, correlation coefficients were computed. The results are summarized in tables 1 and 2. The variables correlated are shown by the subscripts to r according to the following scheme: $1=\mathbf{V}\cdot\nabla T$, $2=\partial T/\partial t$, $3=w$, $4=h$.

The largest coefficients are clearly those between advection and vertical velocity. The positive value means here that cold air sinks and warm air rises; i.e., a direct circulation. The principal significance of these coefficients is that the local temperature change, which is also strongly correlated with the advection, is not closely related to the vertical motion.

On the basis of the correlation coefficients it becomes possible to suggest a pattern of the relationships between the short-wave troughs and ridges and the corresponding components of the vertical velocity. In the vicinity of the short-wave troughs there tends to be a maximum of cold air advection and local cooling. The advection and

TABLE 1.—*Correlation coefficients computed for the period December 17–22, 1953. Subscripts on r refer to deviations from their mean values of the variables as follows:* $1=\mathbf{V}\cdot\nabla T$, $2=\partial T/\partial t$, $3=w$, *and* $4=h$. *n is the number of points on which the respective coefficients are based.*

	n	r_{12}	r_{13}	r_{14}	r_{23}	r_{24}	r_{34}
All days:							
40°	54	−0.442	−0.816	−0.416	−0.158	0.158	0.170
45°	60	−.549	−.852	−.599	.030	.556	.367
50°	66	−.761	−.767	−.676	.167	.749	.463
55°	66	−.767	−.446	−.739	−.233	.730	.101
All latitudes:							
Dec. 17	41	−.502	−.929	−.916	.146	.486	.840
Dec. 18	41	−.040	−.733	−.020	−.651	.645	−.424
Dec. 19	41	−.576	−.852	−.112	.062	.244	−.020
Dec. 20	41	−.387	−.704	−.422	−.254	.365	.201
Dec. 21	41	−.763	−.569	−.823	−.097	.655	.434
Dec. 22	41	−.696	−.848	−.527	.209	.755	.159
All days and all latitudes	246	−.616	−.765	−.592	−.036	.592	.267

TABLE 2.—*Correlation coefficients computed for the period January 1–9, 1956. Subscripts on r refer to deviations from their mean values of the variables as follows:* $1=\mathbf{V}\nabla\cdot T$, $2=\partial T/\partial t$, $3=w$, *and* $4=h$. *n is the number of points on which the respective coefficients are based*

	n	r_{12}	r_{13}	r_{14}	r_{23}	r_{24}	r_{34}
All days:							
40°	81	−0.727	−0.906	−0.537	0.368	−0.509	0.430
45°	99	−.682	−.908	−.324	.314	−.380	.153
50°	90	−.639	−.859	−.285	.156	−.276	.024
55°	72	−.715	−.859	−.549	.256	−.399	.440
All latitudes:							
Jan. 1	38	−.662	−.957	−.331	.414	.722	.122
Jan. 2	38	−.763	−.943	−.435	.504	.483	.494
Jan. 3	38	−.839	−.833	−.609	.397	.673	.233
Jan. 4	38	−.684	−.583	−.199	−.194	.271	.000
Jan. 5	38	−.158	−.905	−.377	−.277	.153	.301
Jan. 6	38	−.642	−.954	−.276	.383	.684	.066
Jan. 7	38	−.758	−.920	+.133	.441	−.358	.010
Jan. 8	38	−.796	−.678	−.303	.094	.080	.402
Jan. 9	38	−.165	−.926	−.550	−.218	.254	.447
All days and all latitudes	342	−.684	−.882	−.394	.259	.353	.259

the local temperature change, being of opposite sign, tend to cancel, and the vertical velocity associated with the short wave is consequently of small magnitude. However, since the magnitude of the advection in general exceeds that of the local temperature change, there is a tendency for the air to sink as it passes through troughs. Whether the vertical velocity which is here associated with the migratory disturbances is sufficiently persistent, or of great enough magnitude, to be of synoptic significance is doubtful. However, in spite of the small magnitude of the correlation coefficients, r_{34}, the agreement between the two independent periods lends credence to the pattern presented above.

In principle, variables which are poorly correlated may still be highly dependent upon one another, if the variables are 90° out of phase. Additional study of the data, however, indicates that this is not the case with regard to the relationship between vertical velocity and contour height.

6. SUMMARY

The vertical velocities in the lower stratosphere do not present the consistent pattern which is normally found in the upper troposphere. Flow from ridge to trough sometimes has a positive and at other times a negative component of vertical velocity. Nevertheless, the patterns of horizontal convergence in the lower stratosphere appear to be continuous with those in the upper troposphere.

The stationary and transient waves have been treated separately. The present evidence supports the findings of Kochanski [3] that, in association with the long-wavelength, quasi-stationary waves, the air subsides in flowing from cold ridges to warm troughs. Study of the shorter-wavelength, transient waves indicates a considerable degree of correlation between temperature advection and the height anomaly (−0.592 and −0.394 in the two independent cases), between local temperature change and height anomaly (+0.592 and +0.353), and between advection and local temperature change (−0.616 and −0.684). There is also a lesser correlation between vertical velocity and height anomaly (+0.267 and +0.293). These results suggest a model in which there is considerable cold advection and local cooling, and a lesser degree of subsidence, in association with the short-wave troughs.

By superimposing the models suggested above, one arrives at a picture of the flow in the lower stratosphere which is in reasonable accord with observations. In particular, it is found that all possible combinations of positive and negative advection, local temperature change, vertical velocity, and meridional flow are to be expected, depending on the relative positions (and presumably also the intensities) of the long- and short-wave disturbances. Nevertheless the vertical velocity will always be well correlated with the advection; that is to say, the effect of local temperature change is small, by comparison, in the short waves, and nonexistent (or almost so) in the long waves. For the short waves the advection and vertical velocity were correlated by −0.765 and −0.882. For many purposes, then, the temperature advection, which can be determined from a single map, can be used as an adequate measure of vertical velocity.

ACKNOWLEDGMENTS

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THE RELATION OF WEATHER FACTORS TO THE YIELD OF WINTER WHEAT IN BOX ELDER COUNTY, UTAH

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ABSTRACT

A multiple linear regression equation is developed which relates the yield of winter wheat to "effective" precipitation during certain months of a crop year. Antecedent precipitation and other weather factors are investigated to determine their effect on yields.

1. INTRODUCTION

The relation of weather factors to crop yields is being studied now more than ever before. The economist is interested in learning more about these relationships so that future economic conditions can be foretold more accurately. Agronomists are concerned with such studies so that, among other things, wiser recommendations for crop planting can be made for various areas. The farmer, the geographer, the crop insurance worker, and others also benefit as our knowledge of weather-crop relationships increases. In this paper an attempt is made to correlate the yield of one of the important dryland crops in Utah, winter wheat, with certain weather factors.

It is a well-known fact that wheat is the most important grain crop in the world. Wheat is adapted to a wide range of climate and soils, but it is grown principally in moderately dry temperate climates having an average annual precipitation of less than 30 inches [4]. The highest average yields are usually obtained in regions with an annual precipitation of 25 to 30 inches [3]; but in Utah the principal dryland farming areas receive less than 15 inches, on the average [9].

There are several weather factors affecting the growth and yield of wheat. Zink [11], in his study of the weather elements important to wheat growth and yield near Nephi, Utah, found the highest correlations were with evaporation, length of drought periods, and rainfall. He also found that correlations with mean temperatures were low, but were higher with maximum and minimum readings. Zink concluded that the best statistical results are obtained by using plant growth stages. However in the present study which is confined to Box Elder County, Utah, little if any phenological data for winter wheat were available, and it was necessary to use weather data for specific calendar periods. Also, there were no evaporation records available for use in this study.

Some of the factors that other investigators have believed to be important are: the amount, distribution, reliability, and effectiveness of rainfall; length of growing season; the amount of sunlight and soil moisture; snowfall; snow cover; and the frequency and duration of very low or very high temperatures [1, 11].

The principal climatic factor limiting the yield of winter wheat over western United States is insufficient precipitation. In the growth of this crop there appears to be at least two critical periods in which adequate soil moisture is essential to the production of maximum yields. One of these is the period during heading, blooming, and filling. Another critical period is the time of planting [8].

Most of the winter wheat in Utah is of the hard red class and is grown principally on "summer fallow" ground. This crop is usually planted in September and harvested in July. Diseases such as rust, scab, mildew, and leaf spots; leaching of plant food from the soil; excessive growth and lodging; and difficulties in planting, harvesting, and caring for the crop are not as much of a problem in Utah as in more humid climates [6].

Although this particular study is confined to Box Elder County, which ranks first in the production of small grains in Utah, it is believed that the results would be applicable to other dryland winter wheat areas of western United States. Box Elder County, in northwestern Utah, comprises an area of 5,600 square miles, with about 90,000 acres planted to dryland winter wheat [7]. Most of the cultivated crops are produced in the southeastern portion of the county. The northern and western sections are used principally for grazing.

2. PROCEDURE AND RESULTS OF STUDY

Winter wheat yield data were obtained from the office of the Agricultural Marketing Service, U.S. Department of Agriculture, Salt Lake City, Utah; temperature and precipitation data were taken from *Climatological Data*, Utah (monthly and annual issues) published by U.S. Weather Bureau.

The first step was to analyse the yield data for time

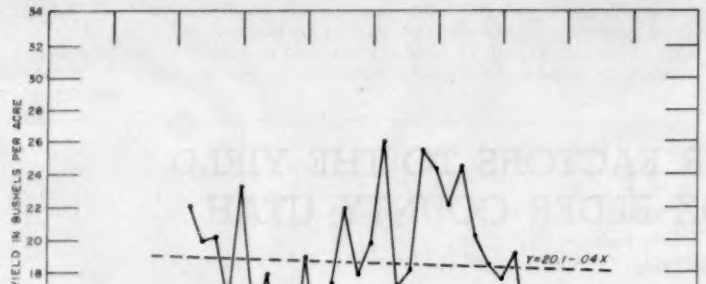

FIGURE 1.—Winter wheat yields, in bushels per acre, for Box Elder County, Utah, plotted against years for the period 1926–55. The regression line is dashed.

trends during the period under consideration (1926–1955) to determine the effect of improved cultural practices, etc., on average yields. According to a study made in 1951 by the U.S. Department of Agriculture [10], the average yield of *all* wheat in the United States showed a pronounced increase during the period 1939–1948, compared to the eight decades prior to 1939. In the Western States, an early peak was reached in 1915, with the yield dropping markedly in 1917 and remaining relatively low through 1940.

In Box Elder County there is no evidence of a marked time trend in the winter wheat yield data from 1926 through 1955. A linear regression line fitted to yields shows a slight decrease of about 1 bushel per acre during this period, but a "t" test indicates this change is not significant at the 5 percent level. The regression line and a plot of annual yields versus years are shown in figure 1.

Precipitation records for ten weather stations in and near Box Elder County were tested for consistency for the period 1926–1955, by the double-mass analysis technique [2]. Of the ten stations, seven were found to have compatible records that could be used without any adjustment of the data. The correlations between seasonal precipitation (July of one year through June of the following year) at five of the stations (Grace, Idaho; Brigham City, Corinne, Richmond, and Riverdale, Utah) and the annual wheat yields in bushels per acre for Box Elder County were computed. From examination of these correlations and some analyses by multiple regression, Grace and Riverdale were dropped and precipitation at Brigham City was assigned a weight of 0.50, Corinne 0.30, and Richmond 0.20.

The effective monthly precipitation for the period under consideration was then computed. Effective precipitation is the sum of the precipitation values times the respective station weights. This procedure is identical with the one outlined by Linsley, et al. [2], except effective precipitation in this study is correlated with wheat yield rather than water-year runoff.

TABLE 1.—*Predicted and actual yields of wheat in bushels per acre for Box Elder County, Utah, for 1956 and 1957*

Year	Predicted yield	Actual yield	Predicted −actual yield
1956	20.3	17.1	3.2
1957	21.4	19.3	2.1

Since it was known that for the production of maximum winter wheat yields, adequate soil moisture is necessary at the time of planting and during heading, blooming, and filling, precipitation data during the months of September, October, May, and June were used in this study. A three-variable correlation was computed, using as parameters the following: Y=annual yield of dryland winter wheat in bushels per acre; X_1=effective precipitation in inches for September plus October; and X_2=effective precipitation for May plus June. The relationship of these variables is shown by the scatter diagram of figure 2. The sloping lines on this chart are least squares lines for annual winter wheat yield. The regression equation was determined to be:

$$Y=8.8+1.67\,X_1+1.85\,X_2,$$

where Y=estimated annual yield of winter wheat. The coefficient of multiple correlation R for this set of data is 0.735. R measures the degree of linear association between the estimated and observed values of the dependent variable [5], in this case between the estimated and observed average wheat yields. R^2, that part of the total sum of squares which has been explained by linear regression, is 0.540. In other words, about 54 percent of the variation in wheat yield can be accounted for by variations in X_1 and X_2.

The average forecast error for the sample was 2.3 bushels per acre. Yields were forecast too low 17 times, with an average error of 2.2 bu. per acre, and were forecast too high 13 times with an average error of 2.5 bu. per acre. The standard error of estimate was computed as 2.73.

Average wheat yields for Box Elder County, Utah, are not available prior to 1926. 30 years of record was considered a minimum amount of data for a study such as this; hence, only 2 years (1956 and 1957) of test data were available. A comparison of the predicted and actual wheat yields for those years is shown in table 1.

3. OTHER FACTORS INVESTIGATED

Winter kill of wheat is occasionally a problem in Utah [9] and, obviously, this reduces the average yield in some years. The amount of damage to winter wheat as a result of low temperatures depends, in part, on the actual values of the minimum temperatures observed and whether there is a snow cover on the ground when the low temperatures

occur. However, the introduction of these variables into the regression equation did not improve the results significantly. (The addition of the lowest winter temperatures at Corinne, Utah, accounted for only an additional 2 percent in the variation of winter wheat yields.)

As was mentioned above, most of the winter wheat in Box Elder County is planted on "summer fallow" ground. One might expect that the inclusion of an antecedent precipitation variable in the regression equation would improve the predictability of winter wheat. However, the addition of the following variables did not change the results significantly: (1) The sum of the effective precipitation amounts for April, May and June of the previous crop year. (2) Effective precipitation for the entire previous crop year.

This is not to say that summer fallowing is not effective. It may be that enough moisture is added to the soil, even in the driest years, so that the amount of precipitation during the fallow year is seldom the principal limiting factor in the production of maximum yields, or possibly the measure of "fallow year" precipitation used does not adequately express the important factor of soil moisture at planting time.

The relationship of other weather factors to the yield of winter wheat was investigated. Some of these were: (1) The lowest monthly winter mean temperature at Corinne, Utah. (2) The mean temperature for the months of April, May, June, and July at Corinne. (3) The effective precipitation for the entire crop year, August through July. (4) Maximum temperature for the months of May and June at Corinne.

4. CONCLUSIONS

Dryland winter wheat yield in Box Elder County, Utah, is related principally to the amount and seasonal distribution of precipitation.

The results of this study indicate the importance of adequate moisture in September, October, May, and June. If sufficient soil moisture is available during those months, there is an excellent chance of an above average yield in the crop year.

The addition of other variables to the regression equation, such as effective precipitation during the previous crop year, winter minimum temperatures, and effective precipitation for the current crop year, did not improve the results significantly.

ACKNOWLEDGMENTS

The author wishes to express appreciation to Mr. Philip Williams, Jr. and Mr. Eugene Peck, both of the Weather Bureau at Salt Lake City, for their interest and suggestions during the preparation of this paper. Thanks also go to Mrs. Dorothy Nelson, who prepared the charts and performed most of the computational work.

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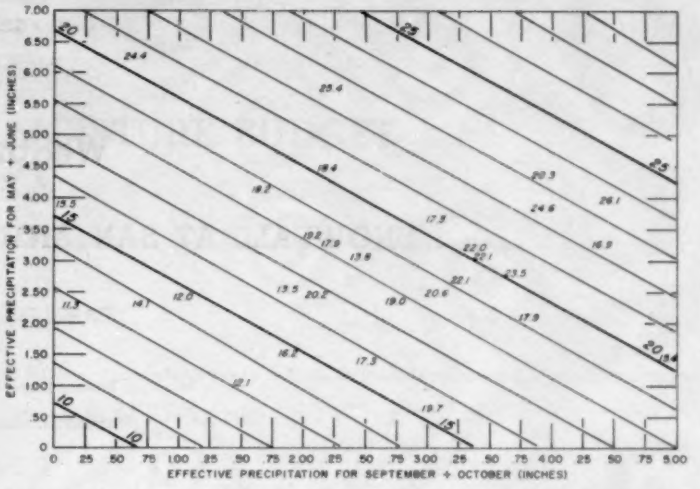

FIGURE 2.—Joint relationship between September plus October precipitation and May plus June precipitation and yield of winter wheat, in bushels per acre, Box Elder County, Utah.

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Weather Note

SNOWFALL AT SAN ANTONIO ON MARCH 16, 1959

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[Manuscript received April 13, 1959]

Snowfall at San Antonio in any month of the year is an unusual occurrence—an event for the local populace. A flake or two of snow falling in San Antonio causes the telephone lines to be jammed at the Weather Bureau, is headlined in the newspapers, and becomes the principal topic of conversation. The total number of days of measurable snowfall at San Antonio from 1885 to 1958, inclusive, is 27, with an average fall of about one-half inch a year. The average for March is a trace.

On the morning of March 16, 1959 hundreds of people called the San Antonio Weather Bureau office to report the falling of snow. A quick look at the meteorological conditions normally would have suggested discounting any reports of snow, as the current temperature was 50° F., and the freezing level was about 6,000 feet above the surface. Yet from 0755 to 0830 CST on March 16 flakes of snow were observed and recorded as a trace at the Weather Bureau and corroborated by a number of experienced observers. The ceiling at the time of the snowfall was 9,000 feet, the lowest reported at any time that day. San Antonio raobs indicated the freezing level at 8,000 ft. MSL at 0548 CST and at 6,200 ft. MSL at 1737 CST. A perusal of all available data indicated that the freezing level was at least 6,000 feet above the surface at the time of the snowfall and the surface temperature never fell below 50° F.

Conditions aloft were favorable for the occurrence of precipitation. At 700 mb. a trough passed over the station. At 500 mb. there was evidence of "digging" to the southwest of San Antonio with a major trough shearing over North Texas, and the forecast cut-off Low in West Texas was evident on the 1200 GMT 500-mb. chart for March 17. The temperature at 500 mb. dropped from −16° C. to −19° C. in the 24 hours ending at 0000 GMT, March 17. The 250-mb. chart for 1200 GMT, March 16 showed two SW-NE jet streams, one over North Texas and one over South Texas. The chart 12 hours later showed one jet stream through South Texas, and later charts showed a continued southward shift of the jet stream. A normal sequence of events followed—a wave was induced and developed on the cold front in the Gulf of Mexico.

From thermodynamical considerations formation of snow aloft in this case was logical, but for it to reach the ground under the conditions on the morning of March 16 does not appear to be so reasonable. It is a common occurrence for aircraft to encounter snow in clouds above the freezing level, and for no snow to reach a surface with temperature well above freezing and far below the freezing level. When snow falling through the atmosphere melts, it absorbs from the air the same amount of heat as was released when it was being crystallized. Certainly, with very little snow falling through a thick layer of air with temperatures well above freezing, there should be no difficulty in the snow melting before it reaches the ground. In fact, it would seem impossible for snow under such conditions to reach the ground.

The most plausible explanation of the March 16 snowfall at San Antonio appears to be that the snow reached the ground in the cool air of vertical downdrafts from high-level thunderstorms or cumulonimbus clouds. There was some evidence of this from a slight increase in the surface winds and an increase in the surface pressure at the time of snowfall. Pilot reports indicated that a high-level instability line moved concurrently with the upper trough. Radar and pilot reports indicated tops of clouds to 31,000 feet. There were a number of pilot reports of snow, sleet, freezing rain, and hail at 4,000 to 8,000 feet MSL east of San Antonio as the trough and high-level line of thunderheads or thunderstorms continued their eastward movement. Beeville reported sleet and rain at 0900 CST; Victoria reported sleet at 0915 CST; and Houston reported small hail at 1010 CST. Before the snowfall at San Antonio and before sunrise, snow was reported in West Texas and eastern New Mexico.

A MODULATED SOIL MOISTURE BUDGET

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ABSTRACT

Most meteorological soil moisture budgets do not account for soil moisture stress changes in the drying cycle or for changes in ground cover or expanding root system. A simple modulated technique is described which considers these factors. Soil moisture stress determined by this technique has a significantly higher correlation with wheat yield than does moisture stress determined by a common method.

1. INTRODUCTION

An important phase of research in agricultural meteorology and soil physics is concerned with soil moisture and the process of evaporation. The soil, plant, and atmosphere each play important roles in this process but are only parts of a continuous system of transport of water from the earth to the atmosphere. While much emphasis has been placed on the individual components of this continuum, there has been less attention devoted to the system as a whole than it would appear to merit.

It is recognized that moisture use by crops cannot pass a maximum which is determined by the amount of energy available, from the sun and the horizontal transfer of heat, to convert liquid water to water vapor. As long as water is freely available for evaporation at the earth-atmosphere interface, the rate of evaporation is mostly dependent upon meteorological parameters. The amount of evapotranspiration required to satisfy the atmospheric demand has been called potential evapotranspiration (PE) [20]. However, as the soil dries, the available moisture decreases, hydraulic tension increases, and transport of water to the interface is insufficient to meet the atmospheric demand and actual evapotranspiration (AE) falls short of PE [6, 7].

The commonest soil moisture budget consists of subtracting daily PE from daily rainfall. This amount is then subtracted (or added when rain exceeds PE) from the moisture present in the soil to give the new soil moisture storage, until available stores have been exhausted. For budgeting irrigation water, this process is continued on a daily basis until a certain permissible soil moisture deficit occurs. At this time an amount of water equal to the deficit is applied to return soil moisture storage to field capacity. This is a simplified description of moisture use by crops under irrigation. "Dryland" soils present many problems with meteorological moisture budgets because they are rarely at field capacity and plant roots explore deep moisture stores.

Most methods of determining PE are semi-empirical [2, 6, 10, 14, 15, 20, 21]. Evaporation pans or atmometers have also been used [3, 5, 8, 12, 24]. As stated previously, when the soil dries out AE becomes less than PE. Thornthwaite suggests that AE is in ratio to the soil moisture in storage (SM). That is, when SM is ½ the total storage possible, AE is ½ PE. He proposes that evapotranspiration continues to near oven dryness. Thornthwaite's data supporting his soil moisture depletion curve were obtained from vapor pressure and temperature profile measurements taken at O'Neil, Nebr. [21]. Curves C and D, figure 1, represent Thornthwaite's concept; curve D continues to the wilting point and C to oven dryness.

Blaney and Criddle [2] correlated actual measurements of consumptive use (evapotranspiration) with monthly mean temperature and daylength in an attempt to obtain a soil-plant coefficient that, when incorporated into their empirical relationship, would give it a wide application. The plant coefficients varied from 0.85 for alfalfa to 1.2 for rice, and applied to semiarid and arid conditions.

Pierce [16] used Thornthwaite's method of calculating PE and compared these data to total water use by 2d year meadow as measured by weighing lysimeters. Thornthwaite's PE was adjusted upward to correct for crop density, age, and rooting depth. Pierce's "dryness" correction curve is represented in curve B figure 1, while his seasonal correction for 2d year meadow (state of growth and time of year) is represented in figure 2. This curve is thought to apply to many soils, provided drainage is not restricted.

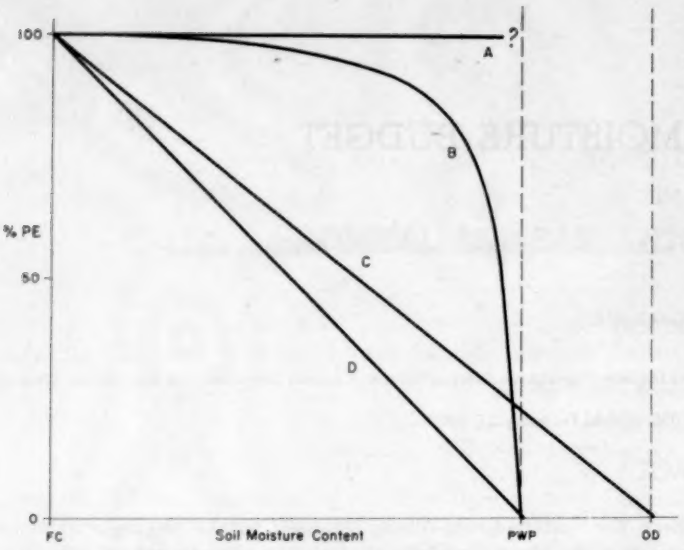

FIGURE 1.—Various proposals for adjustment of PE as soil moisture decreases. A, Veihmeyer; B, Pierce; C and D, Thornthwaite.

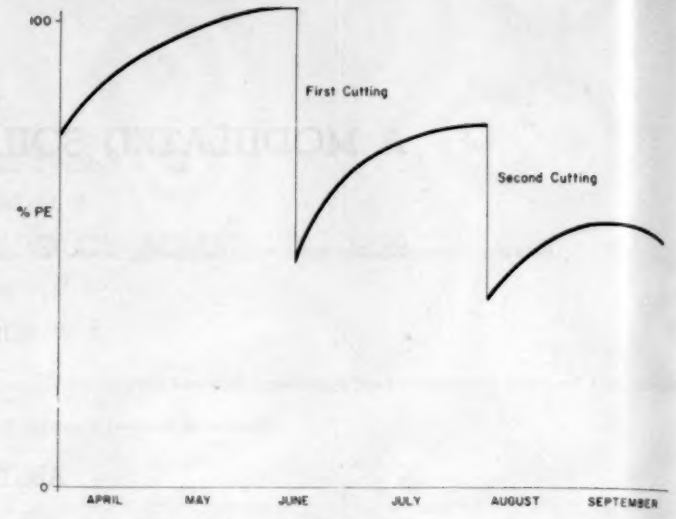

FIGURE 2.—Adjustment of PE for stage of growth of meadow. (Pierce)

Van Bavel's [22] method of determining PE is a simplification of Penman's equation. Nomograms have been drawn to simplify the estimation of daily PE. No allowance is made for soil dryness and crop density (ground coverage) and only moisture within the root zone is considered. (This amount would increase as the crop develops with increasing depth of root zone.) Curve A figure 1 represents Van Bavel's relationship between soil dryness and PE. This curve is similar to that proposed by Veihmeyer [24].

In a recent review of Russian literature, Lemon [11] presents several soil moisture evaporation curves. They are divided into three portions; (a) evaporation proceeds in accordance with the atmospheric demand, (b) evaporation rate declines rapidly as moisture films to the surface become discontinuous and transfer of moisture to the interface decreases, (c) extremely slow moisture movement is dominated by adsorptive forces at liquid-solid interfaces within the soil. Curves in figure 3 suggest this concept. Except for the initial plateau, this is characteristic of many "tension-moisture content" curves.

Marlatt [13] investigated the change in AE as the soil in lysimeters and field plots dried out. By regular soil sampling at 3-inch intervals to 48 inches under a corn crop throughout the season, curves similar to those of Lemon [11] were obtained. He found that AE proceeded at the potential rate up to a point depending chiefly on rooting depth, then AE fell off sharply. These data are represented by the curves in figure 3. The deflection points correspond to the various crop rooting depths at different periods during the season. Philip [17] obtained the same type of curves by analysis of the mathematical and physical aspects of soil moisture evaporation. The horizontal portion of the curves may roughly correspond to the zone of "complete depletion," described by Hagan [4]. This zone may be defined as the volume or, in the case of closely spaced plants, the depth from which soil moisture is withdrawn at or near the potential rate until most of the available water has been removed. The soil in this zone would be thoroughly permeated with fine roots. The second and third portions of the curves may correspond to the "transitional zone" of Hagan [4]. This area contains a varying concentration of roots, decreasing with distance from the plant and extending to a considerable depth.

Marlatt [13] compared the measured and computed soil moisture content of lysimeters and field plots. Measurement was accomplished by sampling, and computation was achieved by adjusting Thornthwaite's PE to soil dryness and rooting depth (according to curves in fig. 3). His correlation was 0.997. Using Van Bavel's [22] nomograms he found departures of the computed and measured SM, particularly when the soil was dry.

2. MOISTURE BUDGETS

To be useful, a scheme for the budgeting or control of soil moisture by the meteorological method should have several characteristics. First, the method of characterizing the drying ability of the atmosphere should (a) be convenient and simple to use, (b) integrate into one measurement all the various meteorological factors affecting the evaporating ability of the atmosphere, (c) (instruments) be free of structural components that distort evaporation measurements.

Second, the manipulation of evaporation data should (a) be realistic in its description of natural processes, (b) have accuracy compatible with the use to which it is put, (c) be simple and practical.

Many of the methods presently used to measure evaporation do not fill the above requirements as they are not simple or convenient. Those that are, do not accurately integrate meteorological factors affecting evaporation into one measurement. Many instruments are cumbersome and are not free of structural disadvantages. A case in point

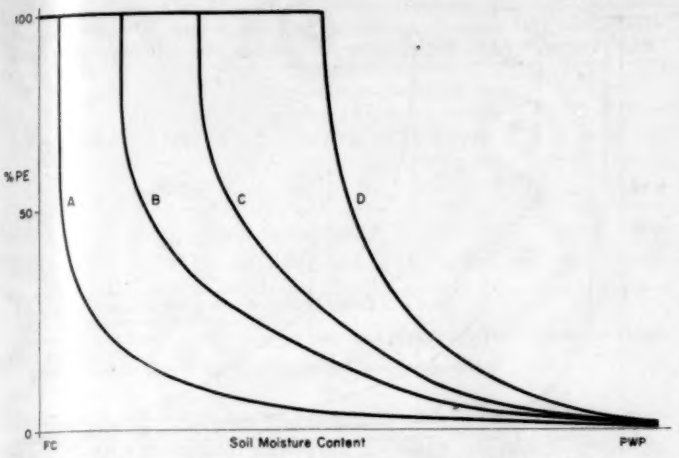

FIGURE 3.—Adjustment of PE for soil dryness and rooting depth of crop (Lemon, Marlatt). Curves A to D correspond to increases in rooting depth of crop.

is that of evaporation pans. Many authors have claimed that atmometers fill most of the above requirements [5, 12, 24], and two recent papers compare an evaporation pan and several atmometers [8, 18]. The black Bellani plate atmometer was described as an accurate and simple instrument that adequately characterizes the evaporation demand of the atmosphere for water.

Similarly, few of our present moisture budget methods (as opposed to evaporation measuring techniques) fulfill all of the above requirements. The accuracy of many present techniques is doubtful; most of them are not realistic in that they do not account for the drying of the soil as moisture loss continues.

It is clear that any accurate description of the moisture status of a soil will not be simple. The plant is the only true indicator of this factor and at the present time it is not possible to measure plant moisture stress, per se. It is necessary then in any budget scheme to make *some* simplifying assumptions, which will be based on crop growth stage, rooting characteristics, and other soil and plant factors. In making these assumptions it is desirable, at the same time, to fulfill the requirements set forth concerning evaporation data manipulation.

A soil moisture budget has been developed which, in the opinion of the authors, fulfills the requirements of the "model" budget described above. Soil dryness is compensated for in a manner similar to that described by Lemon [11], Marlatt [13] and Philip [17]. Crop rooting depth and soil storage may be adjusted periodically during the growing cycle, and rainfall is evapotranspired at the potential rate on the day of the rain and on subsequent days depending on the amount of rain. Percolation, runoff, and soil dryness by zones are also estimated.

The assumptions made in the technique are: (a) all moisture from the zone of "complete depletion," described earlier, is evapotranspired at the potential rate, (b) PE is modulated so that moisture in the "transitional zone" is withdrawn at a decreasing rate depending on the percentage of available moisture remaining, (c) available moisture is withdrawn from the topmost layer of soil, before extraction occurs from lower layers, (d) rainfall occurring after a dry spell is evapotranspired at the potential rate until it has been depleted, unless rainfall percolates into the "transitional zone"; then PE is modulated and withdrawal is at a slower rate, (e) percolation is regarded as complete; if the soil is saturated, rainfall is regarded as runoff. Soil moisture between saturation and field capacity is assumed to percolate.

The weakest assumption is (c) since a plant withdraws water from any zone occupied by roots so long as soil moisture in that zone is available. However, the essentials of the problem are retained since the plant withdraws water from soil zones of varying moisture contents at varying rates, depending on the moisture content of each

TABLE 1.—*Daily PE and soil moisture*

Month—April			Soil moisture storage—Percent of PE											
			100		50		20		10		5			
			Soil zone capacity (inches)											
			0.25		0.25		0.25		0.25		0.25			
Date	PE	Rain	AE	SM	AE	SM	AE	SM	AE	SM	AE	SM	Deficit	Surplus
Last page				0.25		0.25		0.25		0.25		0.25		
1	0.15		0.15	.10										
2	.10		.10	.00										
3	.05	0.06	.05	.01										
4	.08	.05	.06	.00	0.01	.24								
5	.22				.11	.13								
6	.26				.13	.00								
7	.20						0.05	.20						
8	.24						.06	.14						
9	.28						.07	.07						
10	.16						.04	.03						
11	.16						.03	.00	0.00	.25				
12	.10								.01	.24				
13	.05	.35	.05	.25		.05								
14	.03	.33	.03	.25		.25		.10						
Total	2.08	.79	.44	.25	.25	.25	.25	.10	.01	.24	0.00	.25	0.00	0.00
Total AE	.95													

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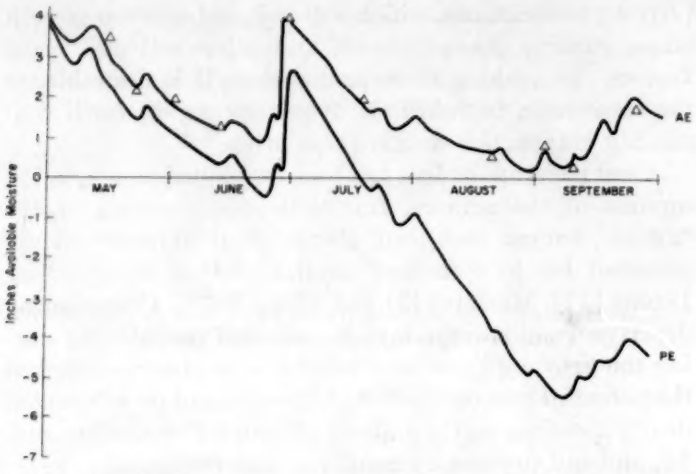

FIGURE 4.—Soil moisture loss calculated by modulated budget (AE) and by simple budget (PE). (Δ indicates spot Coleman moisture block readings.

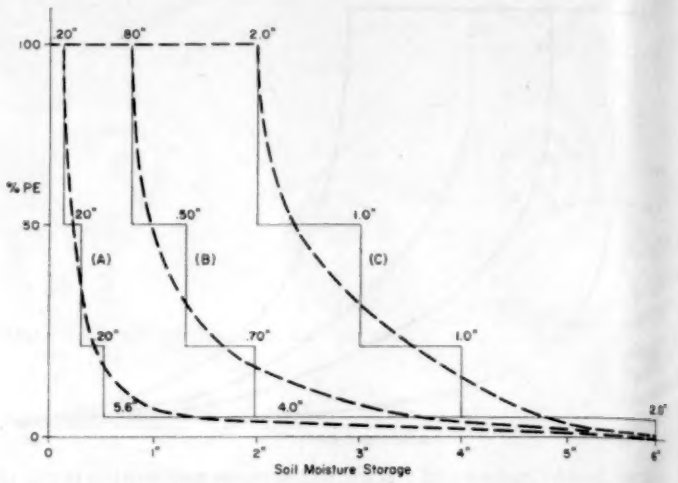

FIGURE 5.—IBM 650 program for adjustment of PE as soil dries and plant roots expand. (Curve A for Aug. 1 to May 31; curve B June 1 to June 30; curve C for July 1 to July 31.)

zone. Recent work by Vasquez and Taylor [24] substantiates this view. There may be some criticism of assumption (e) since local conditions and individual soils may seriously retard percolation.

The individual operations of this technique are set forth in table 1. For purposes of demonstration, the soil zone of complete depletion was assumed to contain 0.25 inch of available water and the transitional zones, 1.0 inch of water, making a total soil storage of 1.25 inches of available moisture. The uppermost 0.25 inch of moisture was evapotranspired at 100 percent of PE; the remaining 1.0 inch was withdrawn in a stepwise fashion with 0.25 inch per step, in such manner that the steps were the best fit to the type of curves in figure 3. In other words, the first 0.25-inch portion of the 1.0 inch in the transitional zone was withdrawn at 50 percent, the second at 20 percent, the third at 10 percent, and the fourth at 5 percent of PE.

Figure 4 compares soil moisture loss by a grass-legume sward calculated by two techniques: (a) the modulated budget described above, and (b) the common type discussed briefly at the beginning of this paper. PE was estimated with a black Bellani plate atmometer [8]. The differences are clearly discernible particularly as the soil begins to dry out. Spot moisture block readings verified the accuracy of the modulated technique.

3. COMPARISON OF TWO BUDGET TECHNIQUES

The two moisture budget techniques partially described above were programmed for an IBM 650 and soil moisture calculations were made for Lethbridge, Alberta, for 36 years (1921–56 inclusive). In both techniques, PE was estimated by Thornthwaite's method [19].

Common Budget (A): The method employed in the simple budget technique has been outlined by Robertson and Holmes [18] and others [20, 22].

Modulated Budget (*B*): The basic characteristics of this technique are outlined in section 2 of this paper. The apportionment of the soil zones and the amount of moisture in each zone is shown in figure 5. The soil drying curve at Lethbridge for stubble or trash covered plowed ground is approximated in the dashed curve A. The stepwise best fit curve used in programming the IBM 650 is drawn over the ideal curve. The length of each step indicates the amount of moisture leaving the soil at the rate of PE shown on the ordinate. This curve was used to calculate soil moisture loss from August 1 (approximate harvest date) to May 31 (seeding date, approximately May 1). The June soil moisture loss curve is represented in curve B, and conditions during July are shown graphically in curve C. These curves were established by laboratory and field observations. Yield data from a 36-year old "continuous spring wheat" rotation at Lethbridge were available for correlation comparisons.

Using the common and modulated budgets, PE, AE, and soil moisture deficits (monthly average below field capacity) during the growth months (May, June, and July) were calculated and compared, through multiple and simple linear correlation, with yields of wheat. The results are shown in table 2.

The data indicate that wheat yields at Lethbridge were not significantly correlated with the evaporating ability of the atmosphere (PE). Soil moisture deficit was more significantly correlated with wheat yields than was actual moisture use (AE). Plant response has been shown to be more closely related to the energy required to take up water (e.g. deficit), than to moisture use per se [4, 13, 17]. The correlation between yield and average monthly moisture deficit below field capacity improves as the season progresses. This indicates that as the crop ages and the roots are more fully ramified throughout the soil, such factors as rooting habit, soil moisture stress, etc., may cause a more pronounced effect on the yield than atmospheric conditions. Allowance is made for these factors in the modified budget (B). Army and Ostle [1] noted an inverse relationship between the evaporating ability of the atmosphere (e.g., PE) and evapotranspiration from

TABLE 2.—*Simple and multiple correlations between soil moisture factors AE, PE, and deficit (calculated by common (A) and modulated (B) methods) and yield of wheat at Lethbridge, Alberta*

Soil moisture	Correlation coefficient and method							
	May r_{y1}		June r_{y2}		July r_{y3}		$r_{y\,123}$	
	A	B	A	B	A	B	A	B
PE[1]	−0.27	−0.27	−0.04	−0.04	−0.14	−0.14	0.28	0.28
AE[2]		.20		.29		**.48		**.58
Deficit	−.37	*−.38	.32	**−.46	*−.44	**−.63	.46	**.68

[1] PE is calculated the same way in both A and B methods.
[2] AE is not calculated in method A.
$r_{y1,2,3}$ etc., indicates correlation between yield and May, June, and July soil moisture factors, respectively.
*, **indicate significance at the 1 and 5 percent levels respectively.

wheat (e.g., AE). An explanation of this anomaly was found in the complex interrelationship of plant growth, available soil moisture, and climate under semi-arid conditions. Results reported here indicate a similar trend. The correlation between August PE and AE is $r=-0.57$ (significant at the 5 percent level).

Even though the results with the modulated budget indicate considerable versatility and improvement over the simpler meteorological methods of estimating soil moisture status, one may wonder why correlations with yield of wheat are not higher. Hopkins [9] pointed out that inhibitory factors such as wind, disease, weeds, insects, etc., are important in many seasons, and that statistical methods may lead to some underestimation of the actual association of yield of wheat and meteorological factors.

4. SUMMARY

The results presented show that a realistic, yet simple meteorological soil moisture budget is possible. Such a budget readily accounts for soil moisture stress and plant rooting characteristics, etc., and yet has accuracy compatible with other methods of determining moisture use by crops. When soil moisture drying curves (such as those in fig. 5), and accurate estimates of the drying ability of the atmosphere such as those obtained with a black Bellani plate atmometer are available, a daily record of soil moisture is possible. By changing the soil drying curves and coefficients, the scheme can be programmed to fit a wide variety of soils, crops, and crop sequences.

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CORRESPONDENCE

Discussion of "A Modulated Soil Moisture Budget"

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January 26, 1959

Holmes and Robertson [1] have demonstrated that mean monthly moisture deficiency is better correlated with wheat yield when it is derived by application of a budget which considers moisture stress and variable root zone than when it is derived from one which neglects these factors. They state that "one might wonder why the correlations with yield of wheat are not higher" and cite Hopkins' view that "inhibitory factors such as wind, disease, weeds, insects, etc., are important in many seasons." The writer feels, however, that one might question the reality of the high correlation found (0.68) in view of the assumed linear effect of mean monthly deficiency. Certainly the yield would approach zero if the deficiency for a brief period of days were sufficiently severe even though the monthly and season values were not critical.

What the authors choose to call a modulated budget is similar in important respects to that employed by the Weather Bureau in connection with its river forecasting program [2, 3]. The differences in the two methods are largely necessitated by virtue of the fact that we maintain the budget for a basin (the order of 100 to 500 sq. mi.), while the authors are concerned with a particular plot or field. Thus while we use a two-layer system, we have not found it necessary to vary the two moisture capacities for changes in root zone.

For the basin, we have and use total runoff (surface and percolation) as an item in the budget, whereas the authors must assume that runoff occurs only when the soil profile is saturated. It should be pointed out that observations of rainfall and runoff show that the concept of a saturation threshold is not sound and results in large errors of estimated daily runoff. Using this concept, one is forced to select unrealistic values of soil moisture capacity if the accounting is to provide the correct evapotranspiration over extended periods.

The authors contend that evaporation pans are "cumbersome and are not free of structural disadvantages" and have purported to show in an earlier publication [4] that the Bellani plate atmometer provides better results than can be obtained from a pan. Their comparative studies have been restricted to a 4-foot sunken pan of the type used in Canada and the conclusions do not necessarily apply to the Class A pan used in the United States which is free of heat exchange with the soil. All known evaporation instruments possess decided deficiencies and the Bellani plate atmometer is no exception. It can be used only during those months when the minimum temperatures are above freezing and it requires repeated calibration to avoid time trends in the data.

We have found [5] that free-water evaporation can be estimated from that of a Class A pan or from certain meteorological factors with about equal reliability. In either case, our moisture budget computations are based on the assumption that free-water evaporation and potential evapotranspiration are equivalent. The computation of potential evapotranspiration from solar radiation, air and dewpoint temperatures, and wind, as well as the moisture budget computations, can be performed on an IBM 650 computer [6].

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Reply

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It has been our experience that while periods of a few days of high moisture stress certainly influence the yield of spring wheat, the effect is not marked unless the deficiency occurs at the time of germination and emergence. According to our budget, the Lethbridge plots were never depleted of available moisture during the entire 36-year period, although the supply was low oftentimes. Any drought periods that affected yield were of sufficient length to be accurately accounted for on a monthly basis. In other words, rainless periods of a week or less were overcome because of moisture held in the low root zones. However, if the drought persisted for longer periods, even this store of moisture became depleted and the effect on the yield was marked.

(*Continued on p. 118*)

A NOMOGRAM TO DETERMINE MONTHLY POTENTIAL EVAPOTRANSPIRATION

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ABSTRACT

A nomogram for the solution of Thornthwaite's empirical relationships among mean monthly temperature, latitude, and potential evapotranspiration is constructed. In the form of alignment charts, the nomogram is simple to use and has the added advantages of the temperature scale being labeled in both degrees Celsius and degrees Fahrenheit and of the potential evapotranspiration scale being labeled in both centimeters and inches.

1. INTRODUCTION

Potential evapotranspiration is defined as the amount of water that would be evaporated from the ground or be transpired by the vegetation if there were no shortage of water at any time during the period of observation. The concept has proven to be a valuable tool in the classification of climates as well as in the computation of water balances [1], [2], [3].

Thornthwaite [4] found an empirical relationship among the mean monthly temperatures, the latitude of the location, and the potential evapotranspiration. The determination of the latter involves the use of two tables and a nomogram and is rather time consuming, especially if series of years for different locations have to be computed.

The nomogram presented here was the result of the desire to have a fast, accurate, and convenient method to determine the potential evapotranspiration according to Thornthwaite's method. The manuscript was ready for publication when the author's attention was drawn to a recent article by Palmer and Havens [5], dealing with the same subject. The alignment charts described in the present paper are easier to use and either alone or in combination with the graphical technique of Palmer and Havens will expedite the computations of potential evapotranspiration.

2. THE THORNTHWAITE FORMULA

The computation of potential evapotranspiration according to Thornthwaite's [4] formula consists of three steps. First, the heat index (J) has to be found for the year and consists of the summation of 12 monthly indices (i), where $i=(t/5)^{1.514}$ and t=the mean temperature of the month in °C. Second, the unadjusted potential evapotranspiration (E) must be computed according to the formula:

$$E=1.6\,(10t/J)^{a} \tag{1}$$

in which

$$a=0.000000675\,J^3-0.0000771\,J^2+0.01792\,J+0.49239. \tag{2}$$

Third, E must be multiplied by a factor for daylight (f), adjusting E to the mean possible duration of sunlight in each month of the year. This factor depends on the latitude of the location.

The heat index (J) can be computed from Thornthwaite's [4] table IV, but it is simpler to construct a conversion scale where both ° F. and ° C. are inserted. Such a scale is presented at the top of figure 1. By adding the i's for each of the twelve months of the year J is determined. This completes the first step.

The second step involves the use of equation (1) which, if we take logarithms and put $a=F(J)$, becomes:

$$\log E-\log 1.6=F(J)\ \log 10t-F(J)\ \log J. \tag{3}$$

This equation can be simplified to:

$$\log E=F(J)\ \log t'-F'(J) \tag{4}$$

by ignoring the constant log 1.6, putting $10t=t'$ and $F(J)$ $\log J=F'(J)$. Equation (4) in determinant form is:

$$\begin{vmatrix} -\log E & 1 & 0 \\ \log t' & 0 & 1 \\ F'(J) & 1 & F(J) \end{vmatrix} =0. \tag{5}$$

Finally the potential evapotranspiration (PE) is calculated according to:

$$Ef=(\mathrm{PE}) \tag{6}$$

which can be written:

$$\begin{vmatrix} -m & km\log E & 1 \\ n & kn\log f & 1 \\ 0 & kmn\log(\mathrm{PE}) & m+n \end{vmatrix} =0 \tag{7}$$

with k, m, and n constants.

3. CONSTRUCTING THE NOMOGRAM

For the purpose of constructing the nomogram the determinant (5) must be multiplied by the non-zero determinant (see [6])

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$$\begin{vmatrix} p_1 & q_1 & r_1 \\ p_2 & q_2 & r_2 \\ p_3 & q_3 & r_3 \end{vmatrix} \neq 0 \qquad (8)$$

which gives:

$$\begin{vmatrix} -p_1 \log E + p_2 & -q_1 \log E + q_2 & -r_1 \log E + r_2 \\ p_1 \log t' + p_3 & q_1 \log t' + q_3 & r_1 \log t' + r_3 \\ p_1 F'(J) + p_2 + p_3 F(J) & q_1 F'(J) + q_2 + q_3 F(J) & r_1 F'(J) + r_2 + r_3 F(J) \end{vmatrix} = 0. \qquad (9)$$

Inspection shows that it is permissible to put $p_2 = p_3 = q_1 = q_3 = r_1 = 0$, but p_1, q_2, r_2, and r_3 must be non-zero. Therefore:

$$\begin{vmatrix} -p_1 \log E & q_2 & r_2 \\ p_1 \log t' & 0 & r_3 \\ p_1 F'(J) & q_2 & r_2 + r_3 F(J) \end{vmatrix} = 0 \qquad (10)$$

in which r_2 may be equal to r_3. By multiplying each row by the reciprocal of its element in the third column, ignoring subscripts, rearranging, and inserting the constant log 1.6, we find the parametric equations:

$$\begin{aligned} y_1 &= (p/r) \log 10t & x_1 &= 0 \\ y_2 &= -(p/r)\ (\log E - \log 1.6) & x_2 &= q/r \\ y_3 &= (p/r) \frac{F'(J)}{1+F(J)} & x_3 &= (q/r) \frac{1}{1+F(J)}. \end{aligned} \qquad (11)$$

The nomogram can now be constructed and consists of two parallel logarithmic scales of equal modulus running in opposite direction and a third line not parallel to these. The slanted line is not straight because $F(J)$ is a polynomial. However, within the range of J's for which Thornthwaite's formula has proven to give adequate results, the line is straight, except for slight curves below $J=10$ and above $J=160$. Below $J=5$ the line approaches zero asymptotically toward $x_3 = (1/1.49239)(q/r)$ since by equation (2), $F(J)=0.49239$ when $J=0$; and above $J=160$ the line curves similarly toward $x_1=0$. Between $J=10$ and $J=160$ the slope of the J line is determined by $t=26.5°$ C. on the t scale and $E=13.5$ cm. on the E scale. At this temperature E is 13.5 cm. regardless of index, For $E=1.6$ cm., equation (3) becomes:

$$F(J) \log 10t = F(J) \log J \qquad (12)$$

which, for $F(J) \neq 0$, gives $\log J = \log 10t$. The t scale can therefore be used to find the J's on the sloping line by connecting 1.6 cm. on the E scale with the different values (in cm.) on the t scale.

For temperatures above 26.5° C. (79.7° F.) values of E are given by Thornthwaite [4]. These are added to the E scale above 13.5 as indicated in figure 1.

A convenient size of nomogram is obtained by making the distance between two vertical scales $q/r=17$ cm. and the length of one logarithmic cycle on the vertical lines $p/r=12.5$ cm.

In order to find the actual potential evapotranspiration a second nomogram is added to the first. Determinant (7) becomes with moduli 12.5 for the E line and 10 for the PE line:

$$\begin{vmatrix} -2.5 & 12.5 \log E & 1 \\ 10 & 50 \log f & 1 \\ 0 & 10 \log \text{PE} & 1 \end{vmatrix} = 0 \qquad (13)$$

so the f line has a 50-cm. cycle, while the distance between the E and PE lines is 2.5 cm. and between the PE and f lines 10 cm. Since the E line now functions only as a pivot line it does not need to be marked. Markings for °F. and inches can be added as desired to the t and PE lines.

4. ACCURACY AND PRACTICAL USE OF THE NOMOGRAM

The nomogram gives results at least as accurate as those obtained by the old method and when used in routine procedures has proven to be much faster, especially if one mean J value is established for any one station and a series of years of record is being processed. The deviations occurring in consequence of the use of a mean J value are negligible.

Speed and accuracy both are increased even more if the nomogram is drawn on "hard board" or similar material. It can then be equipped with a system of pins and slots (for an illustration, see [7]). The J and the pivot line should be grooved. A pin is fastened at the appropriate place on the J line. A slotted transparent arm is made to slide over this pin and is connected via a sliding pin on the pivot line with another transparent arm. When the left arm is placed on the appropriate place on the t line, the pin on the pivot line slides automatically into the right position and the right arm can be placed immediately on the proper place along the f line to obtain the PE.

The f values depend on the month of the year and the latitude of the location. They must be obtained from tables as published by Thornthwaite ([4], p. 93) or Thornthwaite and Mather ([1], p. 98). Although it would be possible to arrange these tables in the form of a graph, it would make a nomogram unnecessarily cluttered and awkward to handle.

This nomogram, as any nomogram, is of principal advantage if series of computations have to be made, such as series of years of monthly PE's at one station or mean monthly PE's at different stations. (In the latter case the stations should be grouped by latitude.)

It is, therefore, convenient to mark the f values belonging to the latitude of the station, or stations, before taking a series of readings. As an example the mean temperature for May in Bridgeton, N.J., is 17.5° C.; $J=58.3$; f, taken from Thornthwaite's table, is 1.23 at latitude 39° N.; and the PE is read to be 9.2 cm., as illustrated in figure 1.

APPENDIX

Table 1, as taken from Thornthwaite [4] presents the f values for latitudes 50° N. to 50° S. for the 12 months of

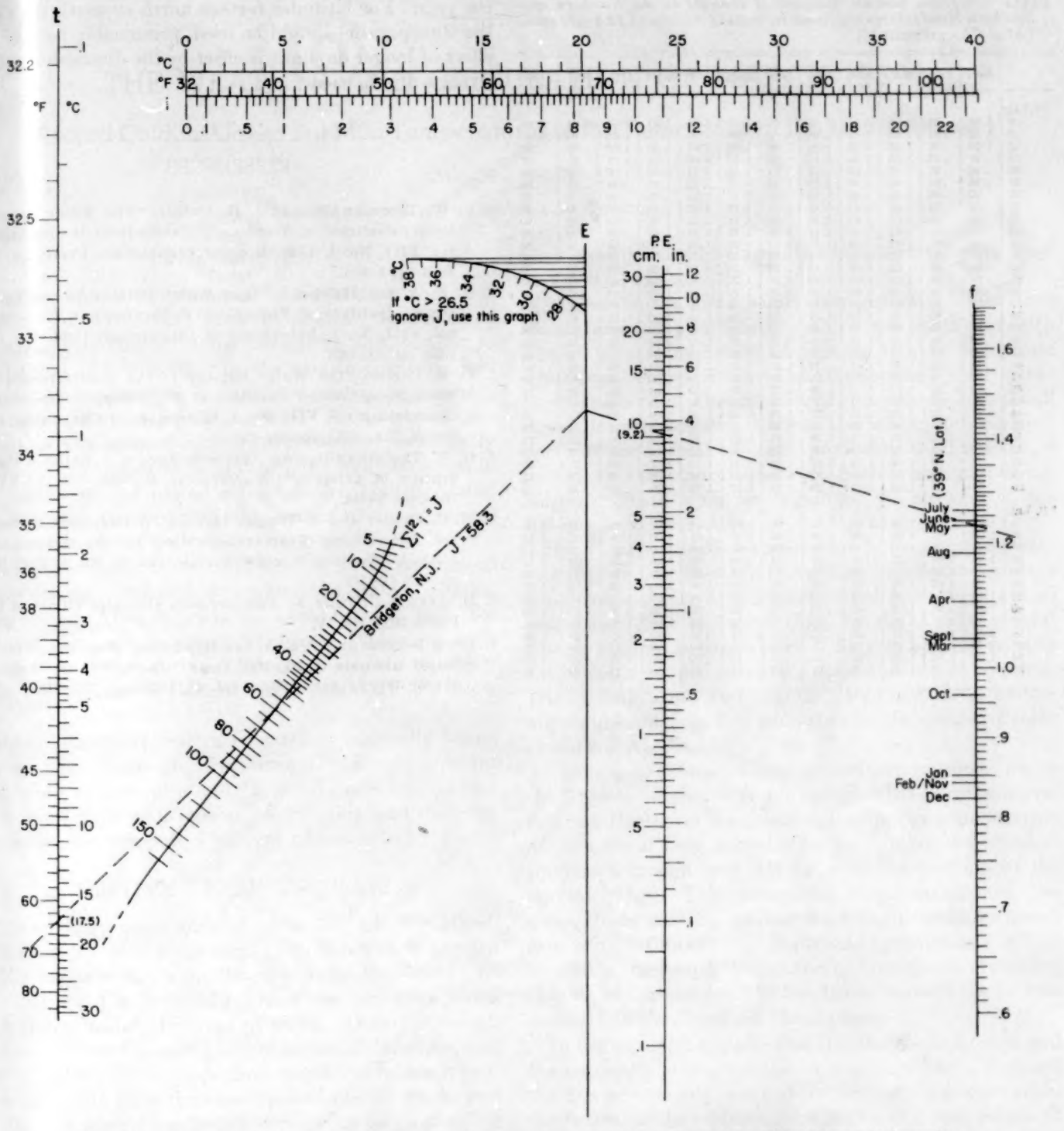

FIGURE 1.—A nomogram to determine the potential evapotranspiration according to Thornthwaite's formula.

TABLE 1.—*Mean possible duration of sunlight in the Northern and Southern Hemispheres expressed in units of 30 days of 12 hours each. (After Thornthwaite [4].)*

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
°N. Lat.												
0	1.04	.94	1.04	1.01	1.04	1.01	1.04	1.04	1.01	1.04	1.01	1.04
5	1.02	.93	1.03	1.02	1.06	1.03	1.06	1.05	1.01	1.03	.99	1.02
10	1.00	.91	1.03	1.03	1.08	1.06	1.08	1.07	1.02	1.02	.98	.99
15	.97	.91	1.03	1.04	1.11	1.08	1.12	1.08	1.02	1.01	.95	.97
20	.95	.90	1.03	1.05	1.13	1.11	1.14	1.11	1.02	1.00	.93	.94
25	.93	.89	1.03	1.06	1.15	1.14	1.17	1.12	1.02	.99	.91	.91
26	.92	.88	1.03	1.06	1.15	1.15	1.17	1.12	1.02	.99	.91	.91
27	.92	.88	1.03	1.07	1.16	1.15	1.18	1.13	1.02	.99	.90	.90
28	.91	.88	1.03	1.07	1.16	1.16	1.18	1.13	1.02	.98	.90	.90
29	.91	.87	1.03	1.07	1.17	1.16	1.19	1.13	1.03	.98	.90	.89
30	.90	.87	1.03	1.08	1.18	1.17	1.20	1.14	1.03	.98	.89	.88
31	.90	.87	1.03	1.08	1.18	1.18	1.20	1.14	1.03	.98	.89	.88
32	.89	.86	1.03	1.08	1.19	1.19	1.21	1.15	1.03	.98	.88	.87
33	.88	.86	1.03	1.09	1.19	1.20	1.22	1.15	1.03	.97	.88	.86
34	.88	.85	1.03	1.09	1.20	1.20	1.22	1.16	1.03	.97	.87	.86
35	.87	.85	1.03	1.09	1.21	1.21	1.23	1.16	1.03	.97	.86	.85
36	.87	.85	1.03	1.10	1.21	1.22	1.24	1.16	1.03	.97	.86	.84
37	.86	.84	1.03	1.10	1.22	1.23	1.25	1.17	1.03	.97	.85	.83
38	.85	.84	1.03	1.10	1.23	1.24	1.25	1.17	1.04	.96	.84	.83
39	.85	.84	1.03	1.11	1.23	1.24	1.26	1.18	1.04	.96	.84	.82
40	.84	.83	1.03	1.11	1.24	1.25	1.27	1.18	1.04	.96	.83	.81
41	.83	.83	1.03	1.11	1.25	1.26	1.27	1.19	1.04	.96	.82	.80
42	.82	.83	1.03	1.12	1.26	1.27	1.28	1.19	1.04	.95	.82	.79
43	.81	.82	1.02	1.12	1.26	1.28	1.29	1.20	1.04	.95	.81	.77
44	.81	.82	1.02	1.13	1.27	1.29	1.30	1.20	1.04	.95	.80	.76
45	.80	.81	1.02	1.13	1.28	1.29	1.31	1.21	1.04	.94	.79	.75
46	.79	.81	1.02	1.13	1.29	1.31	1.32	1.22	1.04	.94	.79	.74
47	.77	.80	1.02	1.14	1.30	1.32	1.33	1.22	1.04	.93	.78	.73
48	.76	.80	1.02	1.14	1.31	1.33	1.34	1.23	1.05	.93	.77	.72
49	.75	.79	1.02	1.14	1.32	1.34	1.35	1.24	1.05	.93	.76	.71
50	.74	.78	1.02	1.15	1.33	1.36	1.37	1.25	1.06	.92	.76	.70
°S. Lat.												
5	1.06	.95	1.04	1.00	1.02	.99	1.02	1.03	1.00	1.05	1.03	1.06
10	1.08	.97	1.05	.99	1.01	.96	1.00	1.01	1.00	1.06	1.05	1.10
15	1.12	.98	1.05	.98	.98	.94	.97	1.00	1.00	1.07	1.07	1.12
20	1.14	1.00	1.05	.97	.96	.91	.95	.99	1.00	1.08	1.09	1.15
25	1.17	1.01	1.05	.96	.94	.88	.93	.98	1.00	1.10	1.11	1.18
30	1.20	1.03	1.06	.95	.92	.85	.90	.96	1.00	1.12	1.14	1.21
35	1.23	1.04	1.06	.94	.89	.82	.87	.94	1.00	1.13	1.17	1.25
40	1.27	1.06	1.07	.93	.86	.78	.84	.92	1.00	1.15	1.20	1.29
42	1.28	1.07	1.07	.92	.85	.76	.82	.92	1.00	1.16	1.22	1.31
44	1.30	1.08	1.07	.92	.83	.74	.81	.91	.99	1.17	1.23	1.33
46	1.32	1.10	1.07	.91	.82	.72	.79	.90	.99	1.17	1.25	1.35
48	1.34	1.11	1.08	.90	.80	.70	.76	.89	.99	1.18	1.27	1.37
50	1.37	1.12	1.08	.89	.77	.67	.74	.88	.99	1.19	1.29	1.41

the year. For latitudes farther north or south the f's of the 50th parallel should be used, presumably because the effect of longer daylight is offset by the diminishing effectiveness of radiation.

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THE WEATHER AND CIRCULATION OF MARCH 1959

Record Cold in Alaska But Mild Temperatures in the Remainder of the United States

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1. INTRODUCTION

The departures from normal of circulation and temperature for March 1959 were rather small over most of the contiguous United States. Westerly flow, similar to the normal [6], prevailed in the mid-troposphere, and maritime polar air masses, with their moderating, springtime effects, dominated the weather. However, the precipitation was more abnormal. There were periodically heavy rain and snow falls as vigorous Lows developed in the Southwest and moved across eastern United States. Some regions received 200–300 percent of their normal precipitation.

In sharp contrast to the mild weather in the contiguous States was the bitter cold throughout the State of Alaska. Long-period Alaskan records were broken, and the reported mean temperatures were more typical of midwinter than early spring. For example, the monthly mean temperature at Barrow was a frigid 27° F. below zero.

Meanwhile, the Hawaiian Islands had unusually warm, dry weather. Honolulu, for example, reported a record high mean temperature and little precipitation. There were just 6 days with measurable rainfall, and the total for the month was only 7 percent of normal.

2. THIRTY-DAY MEAN CIRCULATION

The monthly mean circulation at 700 mb. for March 1959 consisted of a large-amplitude, three-wave pattern in high latitudes and a smaller-amplitude, four-wave pattern in middle latitudes (fig. 1). These two wave trains were approximately 180° out of phase. Only the trough in Eurasia extended continuously across all latitudes, and even it had a marked slope from northeast to southwest. Associated with these truncated, out-of-phase troughs and ridges were several confluence zones with accompanying strong winds. The most prominent ones were in the Atlantic and eastern Pacific, where the height anomalies (fig. 1) indicate stronger-than-normal westerly geostrophic winds. In fact, wind speed maxima were located in both of these regions, and mean speeds were more than 8 m.p.s. above normal (fig. 2). The center in the Atlantic was a primary wind maximum, but the one in the eastern Pacific was essentially just an eastward extension of a larger primary wind maximum located in mid-ocean. All of these wind speed maxima were well east of their normal positions (indicated by arrowheads on heavy dashed lines in fig. 2).

In Eurasia the mean winds were generally weak. Residual blocking over Europe, which was associated with a split jet stream, produced a large area of subnormal wind over central Europe but supernormal wind to the north and south. One jet stream axis was depressed and flowed across northern Africa; the other was displaced north of the normal position, producing wind speeds up to 9 m.p.s. above normal between Scandinavia and Spitzbergen. There were no organized jet streams at the 700-mb. level over Asia.

Returning to the 30-day mean circulation over the oceans (fig. 1), we note that the troughs as well as the primary wind maxima were displaced eastward from their normal positions. This eastward shift of the Asiatic coastal trough resulted in above normal heights and abnormally anticyclonic conditions over Japan and eastern Siberia. The cyclonic areas were also displaced eastward, producing significant negative anomalies in the central Pacific Ocean and Alaska.

The largest anomalies were in northern latitudes, where the 700-mb. heights were far enough above normal to result in a High over Kamchatka, the approximate location of the normal Low, and also far enough below normal to produce a trough over Alaska, near the position of the normal ridge. This anomalous situation affected the downstream pattern, so that the trough normally found over eastern Canada was displaced approximately 25° of longitude eastward to southern Greenland, reducing heights in that area to 510 feet below normal, the largest anomaly in the Northern Hemisphere.

In retrospect, it appears that the block over Europe and the unusually strong trough in western Siberia, through the flux of vorticity, molded the well-marked wave train downstream. In addition, the eastward displacement of waves was favored by stronger-than-normal polar westerlies.

In the United States 700-mb. height departures from normal were small (fig. 1). Over the western States positive anomalies were associated with the flat ridge along the coast. In the eastern States there were weak negative anomalies over a broad band, typical of a flat mean trough. The latter feature suggests that cyclonic activity during March was uniformly distributed over the eastern two-thirds of the United States.

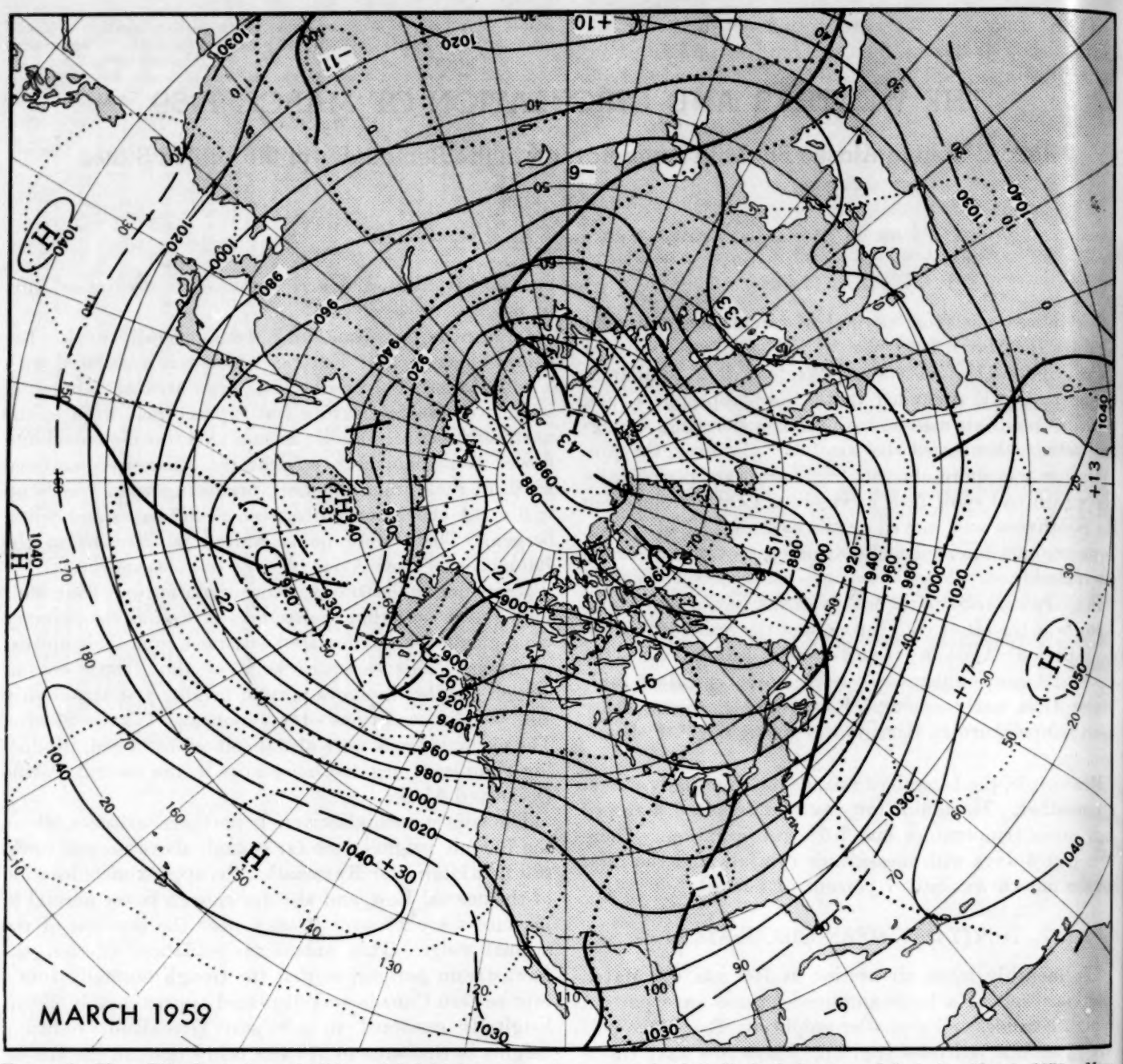


FIGURE 1.—Mean 700-mb. contours (solid) and height departures from normal (dotted), both in tens of feet, for March 1959. Mean ridge in the western and trough in the eastern United States were associated with stormy weather east of the western massif. Abnormally deep trough brought cold weather to Alaska.

3. CHANGE IN MEAN CIRCULATION FROM FEBRUARY TO MARCH

The pronounced block, which was located over western Europe during February [2], was much weaker and farther east during March. This modification of the circulation produced over Great Britain the largest month-to-month height change in the Northern Hemisphere (−610 ft.) and effected over Russia lesser height rises (fig. 3). This was associated with dry weather over Russia and much storminess over Great Britain.

In other areas of the Northern Hemisphere the height changes (fig. 3) were similar to the 700-mb. height departures from normal, which were referred to previously (fig. 1). Hence, in these areas the circulation anomalies for March can be looked upon largely as anomalous height changes from February to March. The rises over eastern Siberia and the eastern Pacific, and the falls over Alaska and central Pacific, all had their counterparts in the height anomaly field. To some extent the same was true in eastern North America, where heights rose in Canada and fell over the United States, resulting in weaker-than-normal west winds or height gradient from the southeastern United States to Hudson Bay.

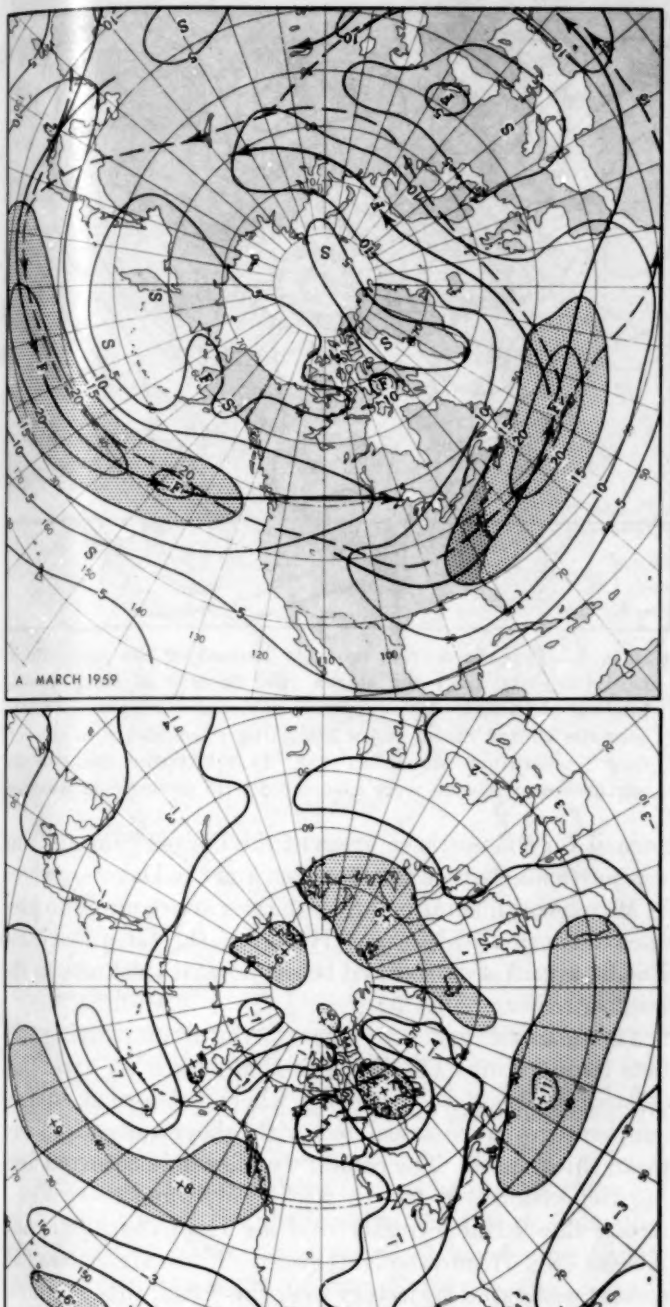


FIGURE 2.—(A) Mean 700-mb. isotachs and (B) departures from monthly normal wind speeds, both in meters per second, for March 1959. Solid arrows in (A) indicate principal axes of maximum winds, and dashed arrows their normal March positions. Regions with wind speeds greater than 15 m.p.s. and anomalies greater than 5 m.p.s. are stippled. Principal wind maxima were located over the oceans east of their normal positions.

4. RECURRENT CIRCULATION AND WEATHER

The 5-day mean flow patterns during March resembled each other and were quite similar to the 30-day mean flow. This is illustrated by the geographical frequencies of

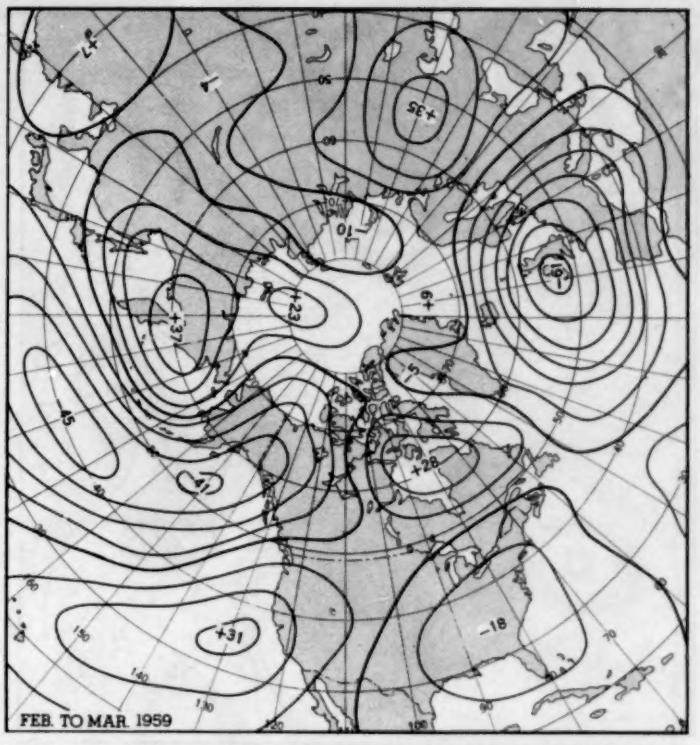
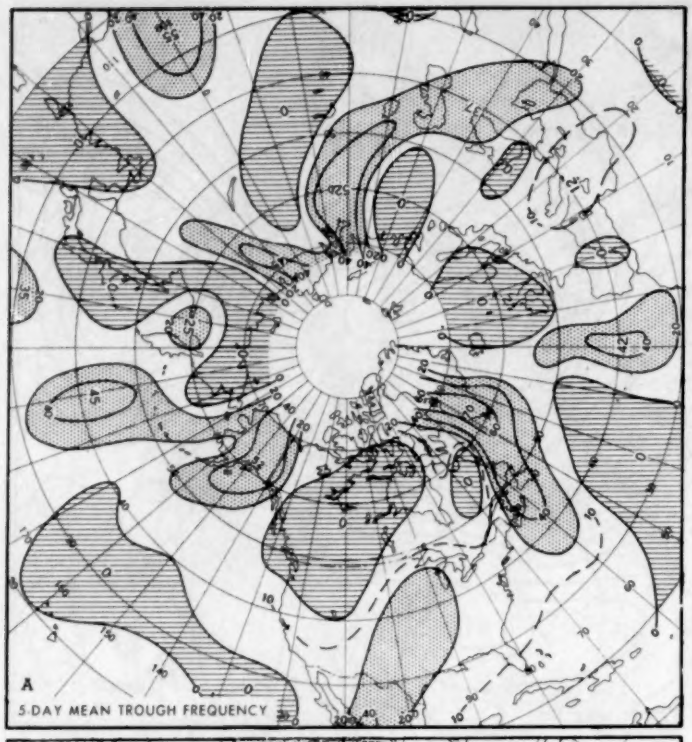


FIGURE 3.—Difference between monthly mean 700-mb. height anomaly for February and March 1959 (March minus February) in tens of feet. Largest falls were over Great Britain. Changes in many areas have striking resemblance to 700-mb. height anomalies for March (fig. 1).

700-mb. 5-day mean troughs and ridges (fig. 4) and by their relationship to the 30-day mean chart (fig. 1). The bands of high frequency of troughs and ridges were narrow, indicating a small variation in the longitudinal position of the troughs and ridges. One exception is the eastern United States, where the 5-day mean troughs were rather uniformly distributed from the Mississippi Valley to the Atlantic Coast.

It is particularly noteworthy that the positions of the 5-day mean troughs and ridges clustered around their respective 30-day mean troughs and ridges. This means that the 30-day mean accurately portrayed the predominant circulation regime of March and was not merely the average of several heterogeneous, short-lived circulations.

The 30-day mean flow of small amplitude across North America was accompanied by zonal trajectories of the Highs for March (Chart IX in [4]). The continental polar anticyclones, which formed in western Canada, glanced east-southeastward and remained almost entirely in Canada. They affected only the northeastern United States where the monthly mean thickness (fig. 5) and surface temperatures (fig. 7) averaged below normal. The remainder of the United States was dominated by maritime Pacific air masses, which also moved eastward across the country. A few anticyclonic centers had trajectories from the Pacific through the Northwest, but many of the Highs first appeared over the Plains and Central States and subsequently moved eastward into the Atlantic Ocean.



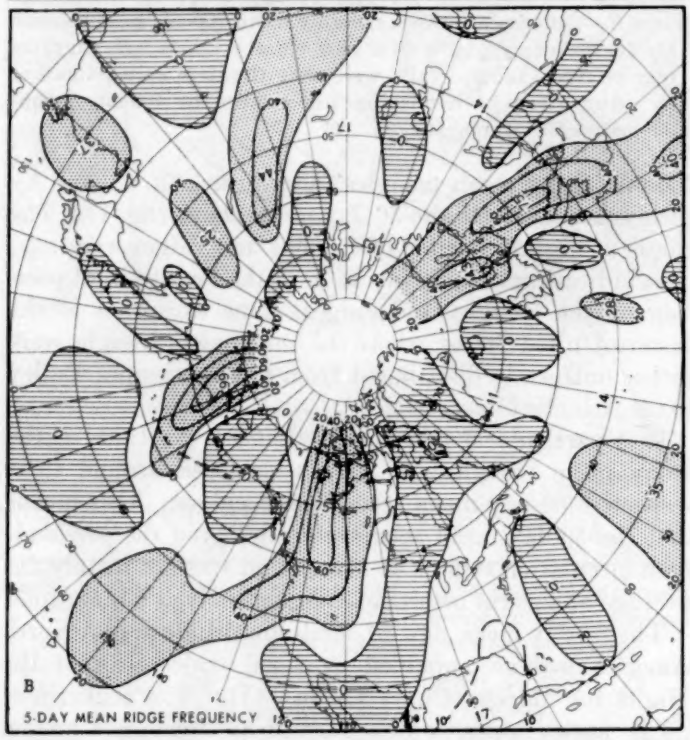


FIGURE 4.—Percent of the time that (A) troughs and (B) ridges on 5-day mean 700-mb. charts were located within 10° longitude intervals at latitudes from 20° N. to 70° N. for March 1959. The data were adjusted to an equivalent basis with 10° at 50° N. as the unit. Isoline interval is 20 percent. Areas with frequency greater than 20 percent are stippled; zero areas are hatched. Note high frequency of troughs and ridges near the locations of the 30-day mean troughs and ridges, respectively (fig. 1).

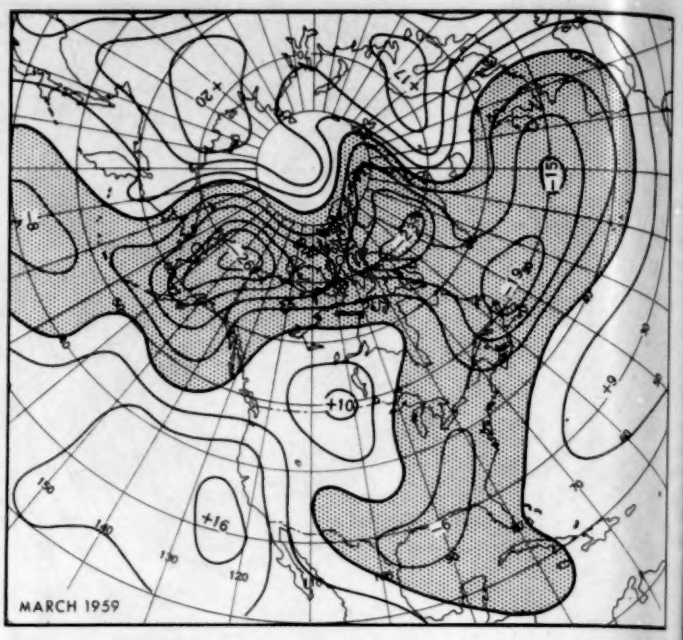


FIGURE 5.—Departure from monthly normal of the mean thickness (700–1000 mb.) for March 1959 in tens of feet. Isoline interval is 50 feet. Below normal values are stippled. Pattern over the United States bears a striking resemblance to the surface temperature anomalies (fig. 7). Extreme negative departures over Alaska were associated with record cold weather.

These maritime air masses had a thickness or mean temperature of the layer from 700 to 1000 mb. that was above normal for the northern areas of the United States. This was particularly true for Montana and the Dakotas, where in March continental polar air masses are required to produce subnormal thickness. However, the same maritime Pacific air masses produced below normal thickness in the southern States (fig. 5).

The same was generally true for the accompanying surface temperatures (fig. 7), but this month an exception existed in the Far Northwest where subnormal surface temperatures were associated with above normal 30-day mean thicknesses. The coolest weather here occurred during the second and fourth weeks [7], when there was a strong flow of abnormally cold air from the Arctic and Alaska (fig. 5) into the Northwest. These air masses had a short and rapid trajectory over the ocean, thus minimizing the diabatic heating, so that the air was still cool when it reached the northwestern United States. During the fourth week the cooling was further enhanced by cyclonic flow aloft.

Along the California coast the weather was unusually warm. Both San Francisco and San Diego, which have long periods of record, reported new high mean temperatures for March. Furthermore, temperatures averaged above normal every day of the month at San Diego and were below normal only one day out of the month at San Francisco. Previously, several authors [1, 3] of this series of articles have related the persistently above normal temperatures in recent years along the California coast to abnormalities of the circulation and of sea-surface temperature. This March the record high temperatures

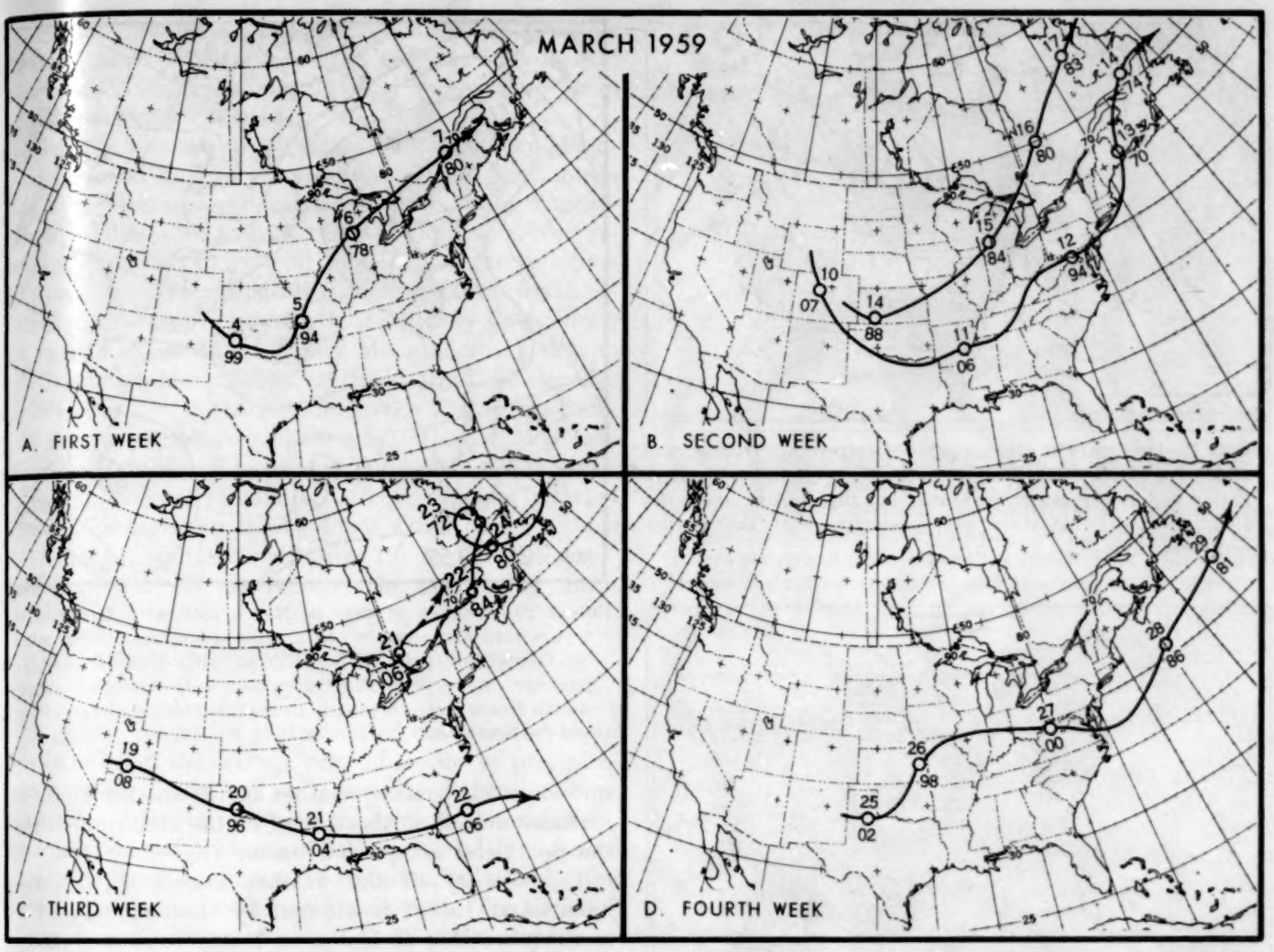


FIGURE 6.—Tracks of selected major cyclones which produced intense storms over the United States during March. Circles are the 1200 GMT locations of cyclone on date indicated by number above position. Number below is central pressure of Low in tens and units of millibars. Each week at least one severe storm moved out of the Southwest producing heavy snow and rain over large areas east of the Continental Divide.

follow directly from the above normal heights at 700 mb., and in addition, from the greater than normal northeasterly flow at 700 mb. and sea level, producing foehn winds for the area south of San Francisco (fig. 1, and Chart XI in [4]).

The cyclones, with two exceptions, which had tracks over the United States also had origins there (Chart X in [4]). Several Lows formed in the Atlantic Seaboard States, but the storms which originated in the Southwest contributed most to the weather over the United States. Each week at least one major cyclone formed over the southern Rocky Mountain States and produced severe weather along its eastward track through the central and eastern United States. A detailed descripition of the effects of these storms is given in [7].

During the first week of March a major storm marked by heavy rains, damaging winds, blowing dust, and drifting snow moved from Colorado northeastward across the Great Lakes into Canada (fig. 6A). The weather during the second week was influenced mainly by two disturbances which reached strong storm intensity as they moved northeastward (fig. 6B). The first storm, 11th to 13th, moved erratically from the lower Mississippi Valley to a position just off the north Atlantic coast and finally turned northward across eastern Maine. This storm produced heavy rains in the South and heavy snows in the Northeast, as high winds raked the entire east coast. More heavy rains in the South and a band of heavy drifting snow from Nebraska to Michigan resulted from the second storm of that week (14th to 16th) which moved from the lower Great Plains across the Great Lakes. During the third week the incidence of Southwest storms continued, but the cyclone which formed over Utah on the 19th moved in an east-southeasterly direction with its path traversing the Gulf Coast States (fig. 6C). In addition, a cyclone formed over Lake Ontario and slowly traced an erratic track across the Maritime Provinces of Canada (fig. 6C), creating stormy conditions in New England on March 21. The mild, sunny weather east of the Rocky Mountains during the first part of the fourth week of March was

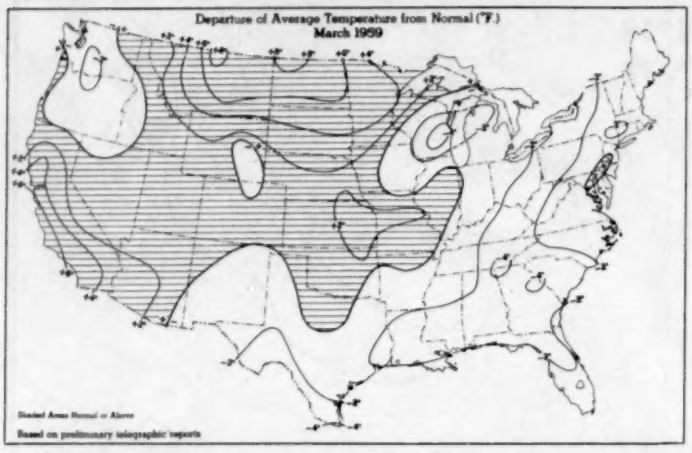


FIGURE 7.—Departure of average surface temperature from normal (° F.) for March 1959. Largest anomalies were in the northern Plains and Rocky Mountain States, but the positive anomalies along the California coast were associated with record temperatures (from [7]).

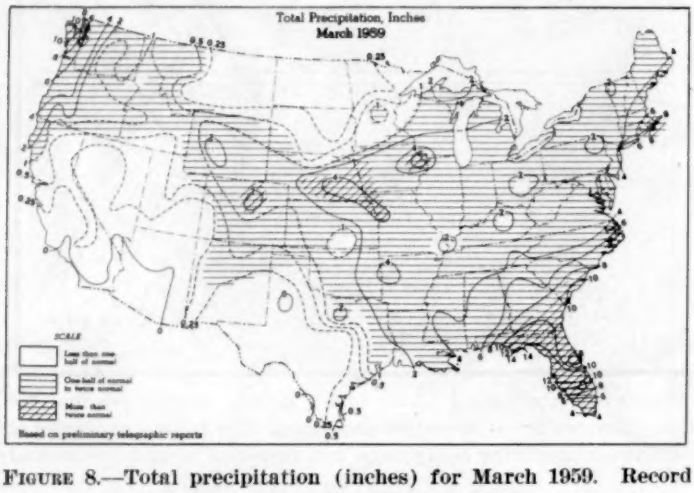


FIGURE 8.—Total precipitation (inches) for March 1959. Record amounts fell in Florida (from [7]).

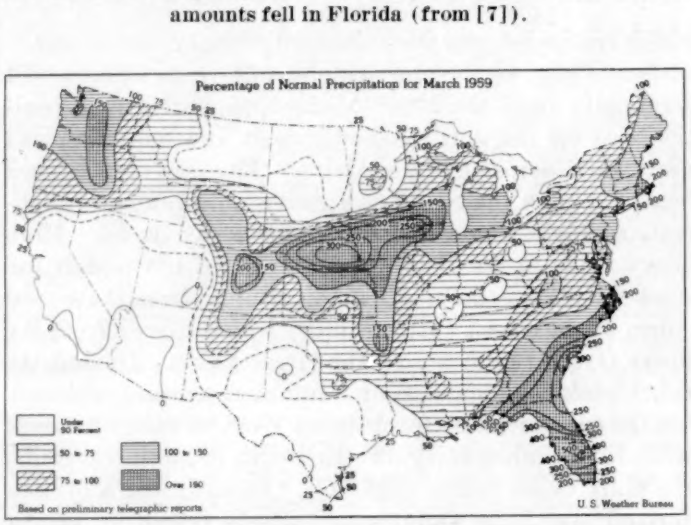


FIGURE 9.—Percentage of normal precipitation for March 1959. March was abnormally wet in the southeastern, northeastern, central, and northwestern States but unusually dry in the northern Plains, northern Rocky Mountains, and the Southwest (from [7]).

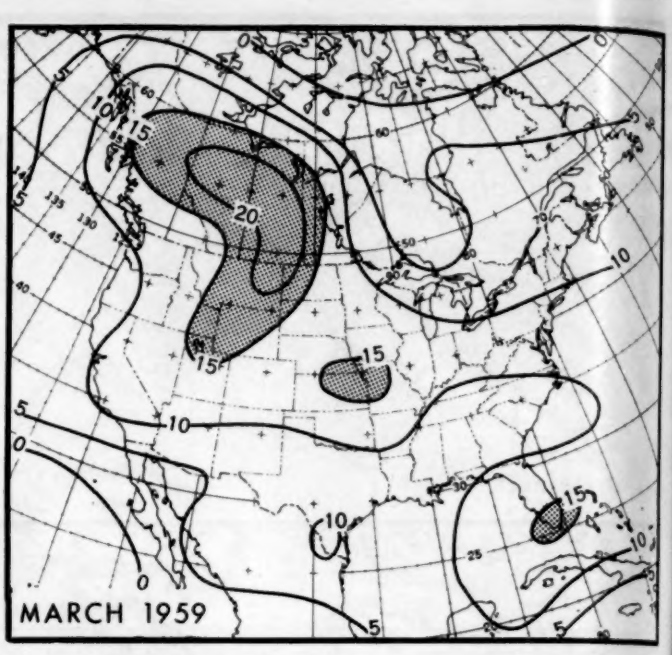


FIGURE 10.—Number of days in March 1959 with fronts of any type within unit squares (with sides approximately 500 miles). All frontal positions are taken from *Daily Weather Map*, 1:00 p.m. EST. Areas with 15 or more days with fronts are stippled. Active fronts were frequently located in Florida, Central Plains, and the Northwest.

interrupted by stormy weather as still another Texas disturbance moved northeastward on the 25th to 27th from the Southwest across New Jersey (fig. 6D).

The severity of the weather associated with these "Southwest Lows" is attested by the summary of the monthly weather at Dubuque, Iowa, which felt the full brunt of several of these storms. It was reported that the precipitation for March was the heaviest since the beginning of records in 1851. The snowstorm of the 4th and 5th brought the greatest amount of snow ever recorded in one 24-hour period for any month. The 6.50 inches of moisture for the month exceeds by 1.48 inches the previous wettest March recorded in 1852. The total monthly snowfall of 30.2 inches is the greatest for any March and the greatest for any month of the year since January 1929. Similar reports of record-breaking weather came from stations in Nebraska and Wisconsin.

The bulk of the precipitation over the United States during March (figs. 8, 9) was produced by the major storms whose tracks are shown in figure 6. However, additional rainfall did occur in the Southeast, associated with fronts which were prevalent over Florida and the Gulf of Mexico (fig. 10). Some of the precipitation was cold-frontal in nature, but the majority was produced by incipient cyclone waves and over-running of the cold air by tropical air masses. Tampa reported the wettest March on record; other Florida stations broke various precipitation records.

In the Northwest above normal precipitation occurred with faster-than-normal west-southwesterly flow across

the mountain ranges (fig. 1), which intensified orographic lifting. Also favorable were the frequent invasions of maritime Pacific fronts (fig. 10) and cyclonic flow aloft. (See *Daily Weather Map* [5].)

That precipitation which fell east of the Rocky Mountains occurred under cyclonically curved 700-mb. mean flow and weak easterly anomalous flow in the Kansas-Missouri-Illinois area; both types of flow have been related empirically to vertical motion and precipitation. However, the explanation for the precipitation which occurred in the easterly flow north of the daily Lows, along the eastern slopes of the Rocky Mountains in Colorado and Wyoming, is not easily extracted from the monthly mean circulation at the 700-mb. level. The vertical motion and precipitation in these central Rocky Mountain States were enhanced by orographic effects not reflected in the upper-level circulation. Also, the precipitation occurred during a few short periods which were not typical of the predominant pattern of the region and consequently would not be reflected in the 30-day mean circulation.

5. RELATION OF PRECIPITATION TO VERTICAL MOTION

Of course, subjective evaluation of the monthly mean flow in terms of the average vertical motion or precipitation is rather crude, but until recently, it has not been practical to objectively estimate the vertical motion. Now, the availability of operational baroclinic models and electronic computers enables us to compute, objectively, the vertical motion for a midtropospheric level. The National Meteorological Center of the U.S. Weather Bureau at Suitland, Md., using a two-level baroclinic model, estimates twice daily the concurrent vertical motion at 600 mb. from initial (observed) data.

The 30-day mean of these values for March 1959 (fig. 11) is a logical pattern roughly consistent with the concomitant mean 700-mb. circulation. In general in figure 11, descending motion (negative values) is found west of the troughs in the northwesterly flow and ascending motion (positive values) east of the troughs in the southerly flow. Exceptions are the negative values just off the northeastern coast of the United States, under and east of the mean trough (fig. 1, 11). This, at least at first thought, does not fit the accepted relationship between horizontal circulation pattern and associated vertical motion, but further inspection of figure 1 reveals that subnormal 700-mb. heights and marked anomalous, northerly flow extended from eastern Canada to New England, suggesting that cold subsiding air masses frequently invaded the area in question, off the northeastern coast of the United States.

Portions of the 30-day mean vertical motion were correlated with the total monthly precipitation (fig. 8). The descending motion over the extreme Southwest and the northern Plains States was related to the driest areas this March. Prescott, Ariz., which had no precipitation, and

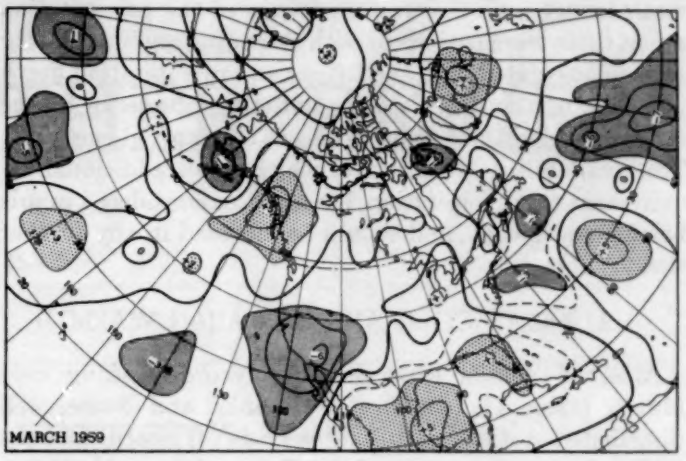


FIGURE 11.—30-day mean vertical motion at the 600-mb. level in millimeters per second for March 1959. Map was obtained by averaging the 60 available twice-daily values computed from baroclinic model and observed data. Isoline interval is 2 millimeters per second. Absolute values greater than 2 are stippled. Descending motion (negative) and ascending motion (positive) tend to be located east of the 30-day ridges and troughs, respectively.

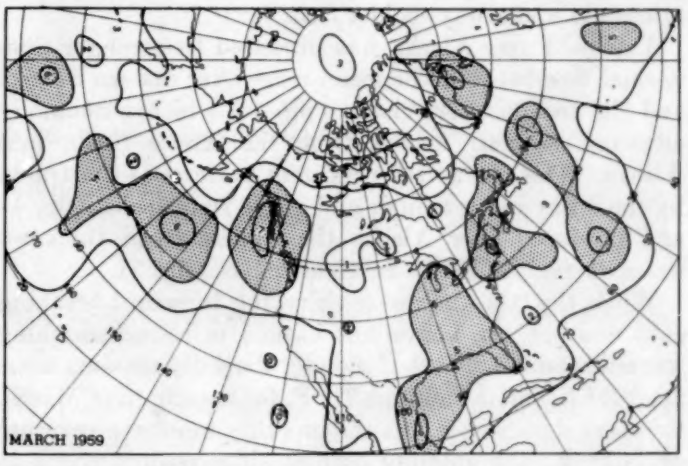


FIGURE 12.—30-day mean of only the ascending (positive) daily vertical motion for March referred to in legend of figure 11. Isoline interval is 2 millimeters per second. Values greater than 4 are stippled. Pattern resembles the total monthly precipitation (fig. 8).

Los Angeles, Calif., reported record dryness for March. Helena, Mont., had the driest March since 1881. The ascending motion over the Far Northwest and the States bordering the Gulf of Mexico fit the precipitation pattern quite well. However, pronounced discrepancies existed in the Nebraska-Kansas-Iowa and Northeast areas where sizable amounts of precipitation occurred in regions of little or no mean ascending motion. This could result from the averaging process where large ascending motions during short periods of storminess would be counteracted by long periods of subsidence. Therefore, the mean of only the positive values (ascending motion) was obtained (fig. 12). As expected, there was considerable ascending motion in these wet regions. Apparently, west of the

mean troughs, on the western fringe of the precipitation areas, there were periods of both ascending and descending motion associated with the lifting and precipitation north and east of the daily storms and the subsidence in the cool air behind these same storms as they moved eastward. The availability of moisture, which is an additional requirement for precipitation, has not been considered in this brief investigation, and should be included in any further study.

6. ALASKAN COLD AND HAWAIIAN WARMTH

Most of Alaska experienced a record-breaking cold March (table 1). Barrow, Fairbanks, and Nome, well distributed stations with long periods of record, had the coldest March on record and Anchorage the second coldest. Departures from normal of the monthly mean temperatures were as large as two or three standard deviations, a rather rare event. Fairbanks, with 16° F. below normal, had the most extreme anomaly, but Barrow, with a frigid mean temperature of 27° F. below zero, suffered the coldest weather in the absolute sense. At the latter city the maximum daily temperature did not exceed −5° F., and the minimum was always colder than −21° F.

This cold over Alaska was produced by stronger than normal flow between the mean ridge over eastern Siberia and the trough over Alaska (fig. 1), which continually advected cold air masses from the Arctic Basin into Alaska. This was a very persistent pattern, as illustrated by the 5-day mean trough and ridge frequencies (fig. 4) and produced over Alaska the coldest mean thickness anomaly for the Northern Hemisphere (fig. 5).

While the Alaskans were shivering in record-breaking cold weather, the Hawaiians basked in record-smashing warmth (table 1). At Honolulu, the daily maxima averaged 79° F. and the minima 71° F. for the month of March. Of more significance was the monthly mean temperature of 74.7° F., the absolute highest since records began in 1905. This month's mean temperature represents a departure of +2.5° F., which is large for this subtropical station that has a standard deviation of only 1.0° F. Similar weather was experienced at Lihue, Kauai, where a record monthly mean temperature of 74.3° F. produced an anomaly of +3.7° F., almost 3 standard deviations above normal.

This warmth was associated with a deficit of precipitation at both stations. Honolulu reported only 0.17 inch, 2.13 inches below normal, and Lihue had 1.37 inches, 2.70 inches less than normal.

TABLE 1.—*Alaskan and Hawaiian surface temperatures (° F.) for March 1959*

Station	Monthly mean	Normal	Anomaly	Standard deviation	Year records start
Alaska					
Anchorage	14.0	24.8	−10.8	4.6	1916
Barrow	*−26.6	−14.9	−11.7	4.0	1921
Fairbanks	*−6.6	9.0	−15.6	6.0	1906
Juneau	33.8	32.8	+1.0	2.8	1915
Nome	*−5.3	8.5	−13.8	6.8	1907
Hawaiian Islands					
Honolulu	*74.7	72.2	+2.5	1.0	1905
Lihue	*74.3	70.6	+3.7	1.3	1905

*Record for March.

The dry and extremely warm weather occurred with a stronger than normal subtropical ridge in the eastern and central Pacific. Over the Hawaiian Islands 700-mb. mean heights (fig. 1), 1,000–700-mb. mean thicknesses, and sea level mean pressures were all above normal during March.

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CORRESPONDENCE

(*Continued from p. 106*)

We were not aware of a similar modulated technique in use by the Weather Bureau.

We are aware that saturation thresholds are somewhat difficult to work with, but for conditions of level plot land, we assume that a saturation point exists and that any precipitation in excess of this value is runoff and hence lost as far as crop use is concerned.

During the past year we have had the opportunity to compare atmometers with the Class A pan, and the buried 4-foot pan. These observations have not caused us to alter our point of view on the disadvantage of pans, generally. Frost is a serious limitation to the use of atmometers. (In this connection we have been experimenting with an instrument which withstands 7 degrees of frost.) Our co-workers in Canada have noticed time trends in Bellani plate atmometers only when air enters the cup or when the surface color or porosity changes because of dirt or accumulated salts.

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